Origin and emplacement of a composite benmoreite - trachyte intrusion: Castle Rock, Lyttelton Volcano

A thesis
submitted in partial fulfilment
of the requirements for the degree
of
Master of Science in Geology
at the
University of Canterbury
by
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University of Canterbury
1991

"...a remarkable dyke forms the crest of the hill, the almost vertical sides of which are in marked contrast with the rounded slopes immediately below it." (Dobson, 1880).

Abstract

Castle Rock and the Eastern Sheet are elongate intrusions associated with the Lyttelton Volcano, Banks Peninsula. These intrusions were emplaced about 9.7 Ma, late in the evolutionary history of the volcano, and are connected to feeder dikes of the Lyttelton radial swarm. Trachytes, Groups A and B, and benmoreites, low silica and high silica, are components of Castle Rock, whereas only Group A trachyte is found within the Eastern Sheet.

The flow foliation and cooling joints of the trachytes and benmoreites indicate the emplacement of bodies with domal cross - sectional shape in a single episode. Velocity gradients in the magmas during emplacement produced groundmass crenulations subsequent to the development of a strong preferred orientation. These crenulations enabled the growing domes to continuously change their shapes, thus facilitating magma movement within. During the emplacement of Castle Rock magma flow was generally away from the magma source, however, restricted magma movement at the northern end during the trachyte intrusion produced magma movement back towards its origin. Overriding of the trachyte and benmoreite magmas at the advancing magma front did not occur during the emplacement of Castle Rock.

Major and trace element geochemistry, mineral chemistry and quantitative modelling indicate that the trachytes and benmoreites were produced by crystal fractionation from a single mafic parent magma. The mafic precursors of the trachytes and benmoreites were derived by partial melting of a depleted mantle source which was enriched in incompatible elements just prior to or during melting.

Multiple populations of feldspar and clinopyroxene phenocrysts within the Group B trachyte and high silica benmoreite and the presence of disequilibrium textures in many of these phenocrysts are interpreted as evidence of magma mixing. Quantitative modelling suggests that magma mixing was a minor process in the magma reservoir and petrographic evidence indicates that it was probably arrested prior to eruption. The Group A trachyte and low silica benmoreite show no evidence of mixing.

The trachytes and benmoreites were fractionated at shallow depths within the crust and erupted following a short residence time in the magma reservoir. Isotopic data suggest that minor amounts of continental crust were assimilated during the evolution of the trachytes and benmoreites. Quantitative modelling indicates that Torlesse Supergroup, or their metamorphosed equivalents, was the probable contaminant.

The trachytes and benmoreites were erupted from a compositionally zoned magma reservoir in which the magmas became less evolved with depth. A fissure, initiated and propagated by a magmatic stress regime developed around a central conduit, tapped the top of the magma reservoir and in time tapped progressively deeper levels. Magma withdrawal was arrested prior to complete emptying of the reservoir.

The emplacement of Castle Rock and the Eastern Sheet occurred at very shallow depths close to or at the surface on the northern flanks of the Lyttelton Volcano.

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1. Introduction

1.1 Introduction

Castle Rock is a composite benmoreite - trachyte intrusion on the northern flanks of the Lyttelton Volcano, Banks Peninsula. This intrusion is an elongate body measuring 250m x 50m in plan, with an exposed vertical extent of 60m. On the eastern side of Castle Rock and associated with it, is a subsidiary trachyte intrusion, hereafter called the Eastern Sheet. Castle Rock and the Eastern Sheet were emplaced during Late Miocene (9.7 Ma) volcanic activity at the Lyttelton volcano, and were fed by dikes associated with the Lyttelton radial swarm.

Recent studies on composite intrusions have revealed information about magmatic processes (Vogel et al, 1987, Higgins, 1990, Mason & Cohen, 1990, Kanaris - Sotiriou & Gibb, 1985). In light of these studies, it was considered that an investigation of the various rock types associated with Castle Rock and the Eastern Sheet could place constraints on magma petrogenesis, magma chamber dynamics and emplacement mechanisms.

A detailed geological, petrographical and geochemical investigation of Castle Rock, and its associated intrusions was conducted with three objectives in mind:

(A) To investigate the spatial and temporal relationships between the intrusive rocks in order to formulate the sequence and timing of the intrusive events.

(B) To investigate the emplacement of the intrusions by studying the macro - and micro - fabrics in the rocks.
(C) To investigate the petrological variation within Castle Rock and the associated bodies and to evaluate the petrogenesis of the magmas.

1.2 Location of the study area

The study area is located on Banks Peninsula (Fig. 1.1) between Lyttelton Harbour and Christchurch City (Fig. 1.2) and is accessed from the east and the west via the Summit Road. At an elevation of 460m at the highest point, the area commands spectacular views of Christchurch City, Lyttelton Harbour, the Southern Alps and the Pacific Ocean.

1.3 Geological setting

Banks Peninsula comprises two Late Miocene stratovolcanoes, Lyttelton and Akaroa, the centres of which are located in Lyttelton and Akaroa Harbours respectively (Fig. 1.2). Both volcanoes have deeply eroded craters and dip slopes that suggest symmetrical domal forms. The original heights of these volcanoes have been estimated at 1500m and 1800m above sea level (Weaver & Smith, 1989), with Akaroa considered the larger of the two.

The geological evolution of Banks Peninsula is summarised by Weaver and Sewell (1986), Sewell et al (1988),

Sewell (1988), Weaver and Smith (1989) and Sewell and Weaver (1990), and only a brief outline will be described herein.

Late Miocene volcanic activity began on Banks Peninsula around 11 Ma and continued for 6 Ma. The volcanic products were erupted from the two main volcanoes and several subsidiary vents onto basement rocks of Triassic Torlesse Supergroup (sandstone, siltstone, mudstone), Cretaceous Mount Somers Volcanics (rhyolite, andesite), Paleocene Charteris Bay Sandstone and Middle Miocene Governors Bay Andesite.

Volcanic activity on Banks Peninsula is divided into four main phases: Lyttelton, Mount Herbert, Akaroa and Diamond Harbour.

The earliest phase (Lyttelton) is associated with the main vents of the Lyttelton Volcano. Hawaiian eruptions, interrupted by brief periods of Strombolian activity, produced lava flows and pyroclastics (Lyttelton Volcanic Group) ranging in composition from basalt to trachyte. Hawaiite was the main rock type produced, with subordinate basalt, mugearite, benmoreite, trachyte and dacite. During the later stages of this phase the SE crater rim was breached. Throughout the history of the Lyttelton Volcano, radial dikes ranging in composition from basalt to trachyte were emplaced. Several of these dikes fed intrusions, including Castle Rock and the Eastern Sheet.

The emplacement of volcanic plugs and eruptions of lava flows from subsidiary vents in the Lyttelton crater floor and on the SE flanks characterise the second phase (Mount Herbert) of volcanic activity on Banks Peninsula. The products of this phase (Mount Herbert Volcanic Group) were

predominantly hawaiite, with subordinate basalt and mugearite. The eruption style was Hawaiian, but phreatomagmatic eruptions occurred during the later stages of volcanic activity.

The third phase of volcanic activity (Akaroa) is associated with lava flows and pyroclastics (Akaroa Volcanic Group) erupted from the Akaroa Volcano. Volcanic activity was centred around the main volcanic vent, however eruptions from subsidiary vents on the flanks did occur. Hawaiite was the main rock type produced during this volcanic phase, with subordinate basalt and mugearite and rare benmoreite and trachyte. Radial dikes, ranging in composition from basalt to trachyte, were emplaced during the activity at the Akaroa Volcano. Several of the trachyte dikes are associated with volcanic domes.

Prior to the commencement of volcanic activity associated with the final phase and subsequent to the activity at Akaroa Volcano, "Church type" lavas were erupted from subsidiary vents on the flanks and floor of the Lyttelton Volcano. The products of this volcanic activity consist of basanitoid, basalt and hawaiite lava flows and a hawaiite dome with related dikes and sill.

Hawaiian style eruptions from numerous subsidiary vents on the flanks of the Lyttelton and Akaroa Volcanoes and in the Lyttelton crater floor occurred during the final volcanic phase (Diamond Harbour) on Banks Peninsula. The deposits produced (Diamond Harbour Volcanic Group) consist of lava flows and volcanic plugs and range in composition from basanite to hawaiite, with rare nephelinite.

The eruption history of the Late Miocene volcanic activity on Banks Peninsula is illustrated on Fig. 1.3.

1.4 Tectonic setting

Prior to the Cretaceous, New Zealand was located on the eastern side of Gondwana in a compressional tectonic regime, which culminated in the Rangitata Orogeny (Bradshaw et al, 1981). The initiation of an extensional tectonic regime in the Late Cretaceous produced a major period of rifting and between 84 - 53 Ma (Johnson & Wellman, 1989) the Tasman Sea opened and New Zealand became separated from Gondwana. This extensional tectonic phase persisted until the Middle Tertiary and was associated with the Late Miocene intraplate continental volcanism on Banks Peninsula.

A change in the direction of the relative motion between the Indo - Australian and Pacific plates in the Middle Tertiary initiated a compressional tectonic phase (Walcott, 1984) which eventually led to the cessation of volcanism on Banks Peninsula. Today, this compressional tectonic phase in New Zealand (Kaikoura Orogeny) is reflected by an eastward dipping convergent margin in the north, separated from a westward dipping convergent margin in the south by a strike slip fault (Alpine Fault) with a compressional component (Fig. 1.4).

1.5 Previous work

1.5.1 Geology

The main emphasis of the early workers on Banks Peninsula was to establish the number of volcanic eruptive phases and to locate the centres from which these phases originated. Haast (1879), Hutton (1885), Marshall (1893) and Speight (1907, 1916, 1924, 1933) all recognised that Lyttelton and Akaroa Harbours had been eruptive centres during the volcanic development of Banks Peninsula. However there were numerous opinions as to the number and location of other eruptive centres that each worker knew must have existed. Comprehensive summaries of developments in the geological ideas of these early workers are contained in Sewell (1985) and Dorsey (1988).

Recent work on Banks Peninsula has involved geological, petrographical, geochemical and isotopic studies on a broad scale, supplemented by detailed investigations of specific areas. These studies have produced a large data base and only work pertinent to the Lyttelton Volcano will be discussed here.

Based on the stratigraphy of Liggett and Gregg (1965), a petrogenetic model for the Lyttelton Volcano was attempted by Price and Taylor (1980). Subsequent to the eruption of the main basalt to trachyte lavas, Price and Taylor (1980) suggested that a late phase of volcanic activity at the Lyttelton Volcano was associated with the eruption of basalt lavas and the emplacement of a trachyte dike swarm. They

concluded that the major and trace element trends, mineral chemistry, REE patterns and quantitative least squares modelling was evidence for the derivation of the main basalt to trachyte trend by crystal fractionation. Sewell (1985) revised the stratigraphy of Banks Peninsula and allocated Price and Taylor's (1980) late phase basalt lavas to the final phase of volcanic activity on Banks Peninsula: Diamond Harbour. This revised stratigraphy negated Price and Taylor's interpretation for the geological evolution of the Lyttelton Volcano, however, their interpretation for the development of the basalt to trachyte trend by crystal fractionation remained valid.

The comprehensive geological, petrographical and geochemical studies by Sewell (1985) on the Lyttelton Volcano established the stratigraphy of the deposits, interpreted the petrogenesis of the magmas, identified the mantle source from which the magmas were derived, and constrained the history of the Lyttelton Volcano to the early phases of Banks Peninsula volcanism. Subsequent studies by Shearer (1986) and Altaye (1989) at specific areas of the Lyttelton Volcano have supplemented this major work. The petrogenesis of the magmas associated with the Lyttelton Volcano is summarised by Weaver and Smith (1989) and only the main conclusions will be briefly outlined herein.

The volcanic rocks of the Lyttelton Volcano belong to the alkaline olivine basalt series. The sodic basalt, hawaiite, mugearite, benmoreite and trachyte are characterised by an alkaline trend in contrast to the dacites which are potassic and subalkaline. Partial melting of a LREE

enriched mantle source produced a parent magma from which the more evolved rocks were derived by crystal fractionation in a large, shallow magma reservoir. The rocks associated with the Lyttelton Volcano have geochemical and isotopic characteristics evident of crustal contamination. Volcanic activity at the Lyttelton Volcano occurred in an intraplate tectonic environment.

1.5.2 Dikes

The main emphasis of previous workers on the radial dike swarm associated with the Lyttelton Volcano was to establish eruptive centres by locating the points of dike convergence.

Marshall (1893) proposed two points of convergence; the first to the south of Quail Island and the second further to the SE.

Speight (1916) established one convergence point located to the south of Quail Island, however in 1938 he reviewed the dike orientations and proposed an eruptive centre between the town of Lyttelton and Quail Island (Fig. 1.5).

Frost (1965) conducted a computer analysis of dike orientations based on data obtained from Speight (1938), and from fifty six dikes in the NW crater rim established a radial centre on Potts Peninsula (Fig. 1.5). The lack of evidence for a volcanic centre at this locality suggests that either the method used in the study was not valid and/or that the limited data on which the method was based influenced the final results.

By studying the convergence of dike trends and the cross cutting relationships between the dikes, Shelley (1987) established a NE migration of two main eruptive centres during the history of the Lyttelton volcano. The earlier centre (Lyttelton 1), located in Head of the Bay (Fig. 1.5), was associated with the emplacement of dikes in the NW and western areas of the crater rim. Shelley (1987) concluded that the dikes in the northern and eastern sections of the crater rim, including the feeder dikes associated with the Castle Rock and Eastern Sheet intrusions, were emplaced during volcanic activity at the later centre (Lyttelton 2) (Fig. 1.5).

The emplacement of the radial dike swarm was considered a late phase event by Speight (1938), and Sewell (1985). However, Weaver and Sewell (1986), Shelley (1987, 1988), Weaver and Smith (1989) and Altaye (1989) all proposed that the radial dikes were emplaced throughout the history of the Lyttelton Volcano.

Haast (1879), Hutton (1885), Marshall (1893), Speight (1907, 1916, 1924), Sewell (1985) and Weaver and Sewell (1986) considered that the radial dikes of the Lyttelton Volcano were predominantly trachytic. This contrasts with Shelley (1988) who proposed that hawaiite was the dominant rock type of the radial dikes with trachyte being subordinate.

1.5.3 Castle Rock

Early workers on Banks Peninsula considered Castle Rock a dike, which varied in composition vertically and horizontally (Dobson, 1880, Speight, 1924, 1938). Two feeder dikes to the south of the Summit Road were considered by Speight (1938) to be linked to Castle Rock, with a third dike to the west an associated arm. Sewell (1985), Weaver and Sewell (1986), Sewell et al (1988), Weaver and Smith (1989) and Sewell and Weaver (1990) recognised Castle Rock as one of six endogenous domes associated with the radial dike swarm of the Lyttelton Volcano.

1.6 Classification

1.6.1 Introduction

In this section, the classification of the rocks is discussed. The Total - Alkalis - Silica (TAS) system (Le Maitre et al, 1989) was used to define rock types. Normative mineral compositions, indicating the degree of silica saturation, is indicated by the appropriate prefix (qz, hy, ne). Sample descriptions are given in Appendix 1 and sample localities are shown on Fig. 1.6. CIPW normative calculations and geochemical analyses are presented in Appendix 4.

1.6.2 TAS classification

A total alkali-silica (TAS) graph with classification fields (Le Maitre et al, 1989) is presented in Fig. 1.7. Each analysis plotted in the TAS system has been recalculated to 100% on a volatile - free basis. In order to remove altered samples, only those with LOI < 2.5wt% were used in the classification. However, the presence of amygdaloidal zeolites suggested exceptions to this criterion and samples 14034 and 14037 have a LOI of 4.01wt% and 3.19wt% respectively.

The rocks located in the trachyandesite (S3) field of the TAS diagram were subdivided into sodic and potassic groups, according to the criterion $Na_2O - 2$ </> K_2O (Fig. 1.8). Fig. 1.9 illustrates that most of these trachyandesites are sodic and are therefore appropriately termed benmoreites. Geochemical, petrographic and mineral chemistry differences within these benmoreites suggest subdivision into two groups. The rocks located near the boundary of the S_3 - T fields will be hereafter called high silica benmoreite. The term low silica benmoreite will be applied to those rocks located near the boundary of the S_2 - S_3 fields. Two benmoreite samples fall in the potassic field. However, it was felt that the presence of zeolitic amygdales within these rocks may indicate the exchange and/or migration of alkalis and adoption of the potassic lithological name, latite, was not judged appropriate.

Rocks falling within the T field are classified as trachyte. Geochemical, petrographic and mineral chemistry

differences within these rocks suggest subdivision into two groups. Group A includes the trachyte in the lower and middle sections of Castle Rock (Areas D, E, F, Fig. 1.6), the Eastern Sheet (Area C, Fig. 1.6) and the sills intruding country rock beneath Castle Rock (Area F, Fig. 1.6). The trachyte in the upper sections of Castle Rock and the intrusions cutting the low silica benmoreite (Area F, Fig. 1.6) are of Group B composition. The occurrence of Group A and Group B trachyte in dike B necessitated subdivision into two segments. Segment B1 (Group A trachyte) includes the section from Trig station EE (452m) to the blunt termination at spot elevation 430m (Fig. 1.6). The Group B trachyte exposed in Area A (Fig. 1.6) belongs to dike segment B2. Discussions of trachyte subdivisions are presented in Chapter 5. The scatter of trachyte samples across the sodic/potassic fields may reflect the mobilisation of alkalis during zeolitisation (Chapter 5). The dike D and dike F samples plotting in the Basalt field of the TAS classification scheme are high - K basalts.

1.6.3 Normative classification

In calculating the norms, total Fe oxide was apportioned using Fe_2O_3 / FeO = 0.3. The calculations were performed using the computer program PETMIN (Eby, 1990). The results of these calculations are illustrated on Fig. 1.10 and indicate that all rock types (trachytes, high silica benmoreite, low

silica benmoreite) have samples with qz, hy and ne normative mineralogies.

The low silica (70%) and high silica (75%) benmoreites are predominantly weakly ne normative (0.56 wt% - 3.80 wt% and 0.29 wt% - 0.86 wt% respectively) with subordinate proportions of the sample population producing hy and qz normative mineralogies.

The Group A trachyte is predominantly (45%) weakly to moderately ne normative with 15% and 28% of the sample population producing qz and hy normative mineralogies respectively. Four samples (13629, 13630, 13636, 13674) from the Castle Rock trachyte contain 2.62 wt% to 0.83 wt% normative acmite indicating a peralkaline composition.

In contrast to the Group A trachyte and benmoreites, the Group B trachyte is predominantly (56%) qz normative with hy and ne normative mineralogies produced by 25% and 18% of the sample population respectively.

To assess the analytical error inherited by the CIPW normative calculations, whole rocks analyses of representative samples (ne and qz normative) were recalculated subsequent to adjusting the SiO_2 by 2 standard deviations (± 0.5 wt%). The ne and qz normative mineralogies for these recalculated analyses (Fig. 1.10) suggest that the analytical error can significantly influence the CIPW normative mineralogy. This is particularly the case because many trachytes appear to be critically saturated with respect to silica.

The effects of zeolitisation can also influence the CIPW normative mineralogy as the addition of alkalies and water

during zeolitisation effectively increases the ne and decreases the hy and qz in the normative calculations.

The TAS - defined terms for the trachytes and benmoreites (Section 1.6.2) are prefixed with the appropriate normative term (Appendix 1). However, when assessing the degree of silica saturation indicated by these normative mineralogies, the effects of zeolitisation and analytical errors on the normative calculations must be considered.

2 Geology

2.1 Introduction

In this chapter, the stratigraphy and structure of the study area are presented and discussed. Section 2.2 describes the country rock into which Castle Rock, the Eastern Sheet and the dikes were intruded. Descriptions of the geology of Castle Rock and the Eastern Sheet are given in Section 2.3 together with an interpretation of the modes of emplacement. Section 2.4 contains geological descriptions of the dikes in the study area and an interpretation of the emplacement of the feeder dikes associated with Castle Rock and the Eastern Sheet. A synthesis of observations and conclusions is presented in Section 2.5.

2.2 Country rock

2.2.1 Introduction

Castle Rock, the Eastern Sheet and the dikes are intruded into country rock which consists of mafic lava flows, lahar deposits and a trachyte sill (Fig. 2.1, back pocket). The geochemical compositions of the country rock lava flows in the following descriptions are taken from Altaye (1989).

2.2.2 Country rock lava flows

The country rock lava flows (Fig. 2.1, back pocket) are aphanitic and porphyritic, grey to black hawaiites and benmoreites, with phenocrysts of plagioclase ± pyroxene ± olivine. Large phenocrysts of amphibole (up to 1cm) and plagioclase (up to 0.7cm) are found in these flows together with rare pyroxenite nodules (up to 5cm). Flow thicknesses range from 1m to 15m. The flow tops are often oxidised and the contact surfaces between flows are usually irregular (Fig. 2.2). Sinuous vesicles (av 1cm), constituting up to 30% of the rock, occur in these lava flows, particularly near the flow tops. These vesicles are often infilled with zeolites. Jointing is either poorly developed or absent, and discontinuous, incipient flow foliation is rare. A knobbly surface typical of aa flows is evident, and spheroidal weathering has produced a characteristic outcrop pattern (Fig. 2.3).

The hawaiite lava flow observed at the northern end of Castle Rock (Fig. 2.1, back pocket) is aphanitic and porphyritic with phenocrysts of plagioclase (av 0.5cm), pyroxene (av 0.3cm) and iddingsitised olivine (av 0.3cm) constituting 20%, 3% and 5% of the rock respectively (Fig. 2.4). This lava flow is predominantly massive, but rare isolated areas are randomly jointed.

2.2.3 Country rock lahar deposits

The poorly sorted country rock lahar deposits (Fig. 2.1, back pocket) are matrix supported and contain clasts (2mm - 1m in length) ranging in composition from basalt to trachyte. Clast shapes are predominantly subangular to angular but subrounded forms are present.

The incipiently indurated to friable matrix is poorly sorted (0.2mm - 2mm). Rare, discontinuous areas of this matrix display parallel stratification accompanied by a crude vertical size sorting in which the particle size decreases upwards (Fig. 2.5).

2.2.4 Country rock trachyte sill

The country rock trachyte sill (Fig. 2.1, back pocket) is an aphanitic, light to medium grey, porphyritic trachyte with phenocrysts of euhedral to subhedral, alkali feldspar (10%) in a fine groundmass. Weathering along the well developed flow foliation has produced a platy structure. Elongate, unfilled vesicles (av 2cm) occur in the sill and often constitute up to 15% of the rock.

2.3 Castle Rock and the Eastern Sheet

2.3.1 Introduction

Castle Rock and the Eastern Sheet are elongate bodies trending 30° east of north, with an eastern offset of 30m occurring one third of the distance along the length of the Sheet. The length and width measurements of 250m x 50m and 300m x 60m for Castle Rock and the Eastern Sheet respectively, produce length : width ratios of 5 : 1 in both bodies.

Trachyte, low silica benmoreite and high silica benmoreite are components of Castle Rock and the Eastern Sheet (Figs. 2.1, 2.1 A, B, C, back pocket). The trachyte was the earliest emplaced component and outcrops in Castle Rock and the Eastern Sheet, whereas low silica benmoreite is restricted to Castle Rock. The high silica benmoreite, which lies between the trachyte and the low silica benmoreite, has the most limited occurrence of the three components, and is restricted in outcrop to the southern end of Castle Rock. The outcropping volume of trachyte ($700,000^3$) is approximately five times greater and one hundred times greater than that of the low silica benmoreite (150,000 m^3) and high silica benmoreite (6000 m^3) respectively.

2.3.2 Trachyte

2.3.2.1 Introduction

The trachyte is an indurated, aphanitic, porphyritic rock, with phenocrysts of alkali feldspar (2%) and ± pyroxene (tr) in a fine groundmass. Fresh samples of the Group A trachyte distal from the country rock contacts, are medium to dark grey in contrast to the Group B trachyte which is green - brown with a silvery sheen.

Zeolite - rich portions, characterised by either a "snowflake" texture (av 0.2cm diameter) or "knobbles" (av 0.5cm diameter) occur throughout the trachyte. At the northern end of Castle Rock, zeolites also infill fine sinuous fractures in the Group B trachyte adjacent to the low silica benmoreite contact.

2.3.2.2 Inclusions

Rare xenoliths of basalt occur in the Castle Rock trachyte adjacent to the country rock contact. These dark grey to black, well indurated, subrounded to angular, xenoliths (up to 4cm) are porphyritic, with phenocrysts (8%) of plagioclase (up to 1mm) in a fine groundmass.

A single, medium grey trachyandesite (Chapter 1) xenolith (15cm) was found in the Castle Rock trachyte adjacent to the country rock contact (Fig. 2.6). This vesicular xenolith is aphanitic and porphyritic with

phenocrysts of plagioclase (av 1.5mm), olivine (av 0.4mm) and pyroxene (av 0.5mm) constituting 1%, 1% and <1% of the rock respectively. Zeolites infill the circular vesicles (av 1mm) which constitute up to 10% of the fine groundmass.

Rare screens of country rock, up to 40cm in width and 2m in length, occur in the Castle Rock trachyte within 1m of the trachyte/country rock contact.

2.3.2.3 Trachyte/country rock contact characteristics

At the sharp, relatively regular, trachyte/country rock contact, the dark red, baked margins (2 - 5cm thick) in the country rock contrast with the cooled margins (5 - 10cm thick) in the trachyte which are characterised by aphyric, green trachyte. The welding of country rock onto the lower surface of the trachyte during the intrusion has produced a "rough" texture on the trachyte contact surface (Fig. 2.7). Adjacent to the contact, the characteristic red colour of the country rock is replaced by a grey - red for distances up to 45cm from the contact.

Contiguous with the lower contact are several apophyses of trachyte (up to 2m x 20cm), which are intruded into the country rock beneath Castle Rock. The apophyses/country rock contacts are sharp and characterised by chilled/baked margins. These apophyses have a sigmoidal configuration along their entire lengths.

The trachyte/country rock contact dips at various angles (11° - 44°) into the trachyte bodies, except at the upper

country rock contact in the Eastern Sheet, where the contact dips NW (Fig. 2.1, back pocket).

2.3.2.4 Intrusions

Two sills, Sill A (15m x 1.3m) and Sill B (12m x 30cm), are intruded into the country rock beneath Castle Rock along the western side (Figs. 2.1A, back pocket, 2.8) and at the northern end (Figs. 2.1B, back pocket) respectively. Distances of 0.5m to 1m occur between these sills and the main trachyte body.

The sill/country rock contacts are sharp and parallel to the main trachyte body/country rock contacts. Dark red, baked margins (1 - 2cm thick) in the country rock at the sill contacts are characterised by a "rough" surface, produced by the welding of the country rock to the trachyte contact surface during intrusion. The rapidly cooled margins (2 - 5cm thick) in the trachyte exhibit aphyric trachyte which grades from a pink colour at the contact to a green. A flow foliation, orientated parallel to the contacts, is developed in both sills. The sill contacts are regular, except at one locality along Sill A, where the upper contact has a marked irregularity. Adjacent to this, elongate, sinuous screens (up to 10cm thick) of country rock are embedded in the trachyte (Fig. 2.8).

The relatively constant width of Sill B (30cm) contrasts with the width of Sill A, which varies from 1.3m to 6cm along the sill length. As the width of Sill A approaches 15cm a

sigmoidal configuration develops, and pieces of trachyte appear to have been broken off, plastically deformed and rotated (Fig. 2.9).

2.3.2.5 Flow foliation

A well developed flow foliation is prevalent in the trachyte of Castle Rock and the Eastern Sheet (Fig. 2.10, back pocket). This flow foliation is crenulated throughout the two bodies, with increasing deformation occurring adjacent to the country rock and benmoreite contacts. As the width of the Eastern Sheet narrows to the north, increasing deformation of the flow foliation is accompanied by an overall sigmoidal configuration.

Measurements of the flow foliation indicate that it gradually changes from horizontal to vertical throughout the two bodies. Sigmoidal fractures, trending at high angles to the crenulated flow foliation have developed at localities where the changes in flow foliation are abrupt (Fig. 2.11). These sigmoidal fractures are absent at localities where the flow foliation trends gradually from horizontal to vertical (Fig. 2.12).

2.3.2.6 Cooling joints

Rectangular, pentagonal and hexagonal columnar joints in the trachyte of Castle Rock and the Eastern Sheet have column

face widths ranging from 10cm to 1m across (av 0.5m). Joints parallel to the column faces break the columns into bands 1cm to 20cm thick. These joints generally follow the flow foliation at localities where the flow foliation is orientated perpendicular to the column axes, but at localities where the flow foliation is not so favourably orientated, the joints cut it at high angles. Column attitudes include inclined, horizontal, vertical and fan shaped.

In the Castle Rock trachyte, the orientation of the columnar joints is approximately perpendicular to the upper and lower contact surfaces. Adjacent to the benmoreite contacts these cooling joints either terminate or exhibit a marked change in inclination. Major cooling joints in the trachyte generally extend across the trachyte/low silica benmoreite contact and continue for distances ranging from 10m above the contact to the top of the low silica benmoreite (Fig. 2.13, A, B, C). The major cooling joints in the trachyte underlying the high silica benmoreite also extend across the contact (Fig. 2.14), but these joints are not as well developed as those extending across the trachyte/low silica benmoreite contact. Minor cooling joints in the trachyte terminate at the upper contacts.

2.3.3 Benmoreites

2.3.3.1 Introduction

The benmoreites are aphanitic and porphyritic with phenocrysts of plagioclase (10%) and pyroxene (3%) in fine groundmasses. Fresh samples of low silica benmoreite are dark grey or blue - black. A blue - grey colour is displayed by the high silica benmoreite.

2.3.3.2 Intrusions

At the southern end of Castle Rock the low silica benmoreite is cut by Group B trachyte and high silica benmoreite dikes (Figs. 2.15, 2.16) and a Group B trachyte sill (Fig. 2.17). Chilled and baked margins are not apparent at these sharp, regular to irregular intrusive contacts. The widths vary along the lengths of the intrusions and range from 10cm - 1m in the dikes and 5cm - 30cm in the sill. Beneath the trachyte sill, a lens of high silica benmoreite (Fig. 2.17) (3m x 10cm) occurs. The incipient flow foliation in the trachyte intrusions is orientated parallel to the contacts and is often crenulated adjacent to the contacts. Irregularly spaced cooling joints are developed in all of the intrusive bodies.

2.3.3.3 Contact characteristics

Contact zones (2cm - 30cm wide) characterise the sharp, regular contacts (Figs. 2.1 A, B, C, back pocket) between the Group B trachyte and the benmoreites at all localities except those at the southern end of Castle Rock. These contact zones exhibit a dark red to yellow, unfoliated, uncrenulated, aphyric trachyte (up to 3cm thick) overlying a friable, green - brown, extensively crenulated, flow foliated trachyte (Fig. 2.18). This crenulated flow foliation is plastered onto the overlying and underlying rock. Subsequent erosion of these contact zones has produced gaps at the contact (Fig. 2.19). An increase in the deformation of the trachyte flow foliation is apparent up to 3m below these contact zones (Fig. 2.14). No evidence of flow foliation displacement is apparent across the trachyte cooling joints adjacent to these contacts (Fig. 2.20).

At the southern end of Castle Rock, the sharp contacts between the Group B trachyte/benmoreites and the high silica benmoreite/low silica benmoreite are not characterised by these contact zones. Baked/chilled margins are absent at these regular to irregular contacts.

2.3.3.4 Cooling joints

The low silica benmoreite is characterised by well developed columnar jointing. These columns are predominantly pentagonal, although rectangular and hexagonal forms are

found in the northern end of Castle Rock. The column face widths (d) range from 20cm to 1m across, with an average size of 0.5m. Joints parallel to the column faces break the columns into bands 1cm to 15cm thick (av 6cm). The column lengths (l) range from 35m at the northern end of Castle Rock to 8m at the southern end of Castle Rock producing aspect ratios (l/d) of 70 and 16 respectively. Many of the low silica benmoreite columns are broken up into smaller columns (10 - 40cm) by cross joints which do not extend into adjacent columns.

At the northern end of Castle Rock, the inclination of these columnar joints reflects the orientation of the related cooling joints in the underlying trachyte (Section 2.3.2.6) and not the lower cooling surface of the benmoreite. The cross joints in the low silica benmoreite do not have corresponding cooling joints in the trachyte and usually terminate approximately 1m above the lower contact.

Column attitudes in the low silica benmoreite include inclined, vertical and fan shaped. The vertical columns are restricted to the northern section of Castle Rock in contrast to the inclined columns which occur along the entire length of the low silica benmoreite outcrop. Fan shaped columns are restricted to the western side of Castle Rock.

The high silica benmoreite exhibits poorly developed columnar jointing which does not extend across the upper contact.

2.3.4 Interpretation of field evidence

The presence of chilled and baked margins at the Group A trachyte/country rock contact implies that the trachyte was emplaced into cold country rock. Welding of country rock to the trachyte during emplacement produced a "rough" texture on the trachyte contact surface.

The trachyte sills beneath Castle Rock were emplaced in the early stages of the intrusion, as the displacement of the overlying trachyte by the intruding sill would be difficult to achieve given the small volume of trachyte in the sills relative to the large volume of overlying trachyte. The relationship between sill width and sill configuration suggests that at narrow sill widths (< 15cm), the ability of the injecting magma to intrude the country rock was reduced.

During the intrusion of sills A and B and the lower sections of Castle Rock, xenoliths and lenses of country rock were picked up and incorporated in the trachyte. The scarcity of these inclusions and their locality adjacent to the country rock contact suggests that the country rock was a reasonably coherent body and that the ability of the trachyte to incorporate country rock may have been reduced by its high viscosity.

The gradual vertical to horizontal changes in the trachyte flow foliation suggest emplacement of a body with domal cross sectional shape in a single episode. Movement within the growing dome was accommodated by crenulation of the flow foliation and at localities where the flow foliation changed abruptly, movement was facilitated by shearing at

high angles to this crenulation. The parallel orientation of the flow foliation to the country rock contact implies that the contact configuration had some influence on the developing flow foliation. The increase in the flow foliation crenulation and the sigmoidal configuration at the northern end of the Eastern Sheet suggests that as the propagating fracture was arrested, the ability of the magma to intrude the country rock was reduced.

The contact characteristics of the three units in Castle Rock suggest that the emplacement of the trachyte, high silica benmoreite and low silica benmoreite were closely followed one by another. However the presence of the sill and dikes intruding the low silica benmoreite at the southern end of Castle Rock implies that a small proportion of Group B trachyte and high silica benmoreite was emplaced during and/or after the low silica benmoreite. The baked margins reflect some cooling of the Group B trachyte prior to the benmoreite emplacement. Apart from the rare fine fractures at the northern end of Castle Rock, the lack of brittle deformation structures in the Group B trachyte adjacent to the benmoreite contacts suggests that significant solidification of the trachyte did not occur prior to the emplacement of the benmoreites.

The absence of flow foliation crenulation displacement across cooling joints in the trachyte adjacent to the benmoreite contacts implies that stresses imposed by the overriding benmoreites were negligible.

At the southern end of Castle Rock, the absence of baked/chilled margins and the evidence of extensive plastic

deformation in the trachyte suggests that benmoreite eruptions followed the emplacement of the trachyte more rapidly at this locality relative to the northern sections of Castle Rock. This may imply a low silica benmoreite source adjacent to the southern end of Castle Rock.

Numerical modelling and field observations suggest that as magma cools and solidifies, cooling joints propagate away from the cooling surfaces (Peck et al, 1977, Jaegar, 1968, Peck & Minakami, 1968, Degaff & Aydin, 1987). These joints may extend down and up into the magma body from the upper and lower surface respectively, producing two joint sets whose axes are generally perpendicular to the cooling surfaces (Peck & Minakami, 1968). However, irregular temperature distributions as the result of convection in the cooling lava may produce columnar joints at angles other than perpendicular to the cooling surface (Spry, 1962). In the Castle Rock trachytes, the perpendicular orientation of the columnar joints to the contacts reflects cooling from an upper and lower surface. The emplacement of the benmoreites closely following that of the trachytes impeded cooling from the trachyte upper surface, thus producing the termination or marked change in column inclination adjacent to the benmoreite contacts. Cooling joints emanating from the lower surface of the benmoreites utilised the planes of weakness established by the cooling joints in the upper surface of the Group B trachyte and this resulted in many of the benmoreite columns inclined at angles other than perpendicular to their lower cooling surfaces. Even though the disintegration of the preferred orientation in the groundmass of the low silica

benmoreite adjacent to the Group B trachyte (Chapter 4) may reflect convection at the base of the low silica benmoreite magma, the absence of extremely distorted flow foliation in the trachytes and benmoreites implies that convection in the magmas was not prevalent at the time of emplacement and cooling.

The columnar joints in the trachytes and low silica benmoreite developed by discrete cracking events involving the propagation of cracks normal to the column axes (DeGraff & Aydin, 1987). These events produce a series of bands orientated parallel to the column face ends. Subsequent to the development of the major columns and bands, further cooling produced cross joints which broke the columns into smaller pieces.

The cooling surfaces reflected by the inclined, horizontal, vertical and fan shaped columnar joints in the trachytes and low silica benmoreite suggest elongate domal bodies. The relatively shallow dipping flow foliation in the trachyte at the northern end of Castle Rock and the evidence presented in Chapter 4 suggests that the surface over which the low silica benmoreite flowed (i.e. the top of the Group B trachyte) was essentially flat. This implies that the fan shaped and inclined columnar joints of the low silica benmoreite represent a domed upper surface.

The magma source for lava flows can be determined by assessing the direction of columnar joint inclination. Following the technique of Waters (1960), where the magma source is assumed to face the dip direction of the inclined columnar joints, an elongate magma source was determined for

the low silica benmoreite from which magma effused in northerly and westerly directions.

Column regularity, i.e. attitude, size and shape, is determined by the viscosity, temperature and homogeneity of the lava and the rate and regularity of cooling (James, 1920). The coarse column sizes of the trachytes and low silica benmoreite reflect a relatively slow cooling rate. The regularity of the size and shape of the columns in the low silica benmoreite contrast with the irregularity of the same column characteristics in the trachytes suggesting that the low silica benmoreite was more homogeneous in viscosity and/or composition and/or temperature than the trachytes. The major geochemical differences between the Group A and Group B trachyte are established in Chapter 5 along with the homogeneity of the chemical composition of the low silica benmoreite.

2.4 Dikes

2.4.1 Introduction

Dikes A, B, C, D, E and F (Fig. 2.1, back pocket) are associated with the radial dike swarm of the Lyttelton Volcano, the centre of which has been placed at various localities in Lyttelton Harbour (Chapter 1). An examination of the convergence of radial dike trends led Shelley (1987) to suggest a NE migration of two centres of activity (from Lyttelton 1 to Lyttelton 2) during the history of the

Lyttelton volcano (Fig. 2.21). Shelley (1987) concluded that the dikes in the N and E sections of the Lyttelton crater rim, including the feeder dikes of Castle Rock and the Eastern Sheet, were emplaced during activity at the later centre (Lyttelton 2).

In order to facilitate the discussion in Section 2.4.8, the exposure of dikes B, C, D and E in the Summit Road will hereafter be called Area A, while the outcrops of trachyte (Group A and Group B) and low silica benmoreite on the northern side of the Summit Road opposite Area A will be hereafter referred to as Area B (Fig. 1.6).

2.4.2 Dike A (low silica benmoreite)

Dike A is a northward trending, sinuous low silica benmoreite dike which varies in width from 2.5m at elevation 460m to 3.5m at elevation 420m (Fig. 2.1, back pocket). The vertical flow foliation trends parallel to the dike margins and cooling joints are developed across the dike width. This aphanitic, light grey rock is porphyritic, with phenocrysts of plagioclase (up to 2mm) and pyroxene (up to 1mm) constituting 1% and <1% of the rock respectively. Circular to elongate vesicles (up to 0.4mm) infilled with zeolites, comprise up to 10% of the fine groundmass.

2.4.3 Dike B (Trachyte)

Dike segments B1 (Group A trachyte) and B2 (Group B trachyte) contain spherulitic zeolites which are characterised by a white "snowflake" texture (average diameter = 1mm). The textures and composition of the trachyte in dike B are similar to those in the Group A and B trachyte described in Section 2.3.2.1 and will not be discussed further.

The vertical, dike segment B1 (Group A trachyte) varies in width from 9m, at elevation 452m to 4m at elevation 430m (Fig. 2.1, back pocket). The 350° strike at the Trig station (452m) trends to 30° east of north along the segment length. The outcrop at 430m is characterised by a blunt termination (Fig. 2.22) and Group A trachyte does not outcrop again in dike B until Area B (Fig. 1.6). Changes in the orientation of the well developed flow foliation across the dike width are apparent at elevation 452m, where the vertical flow foliation, at the dike margins (Fig. 2.23) trends to high angles (40° to 70°) westward (Fig. 2.10, back pocket) in the centre of the dike. Cooling joints cut the flow foliation at right angles adjacent to the margins but at high angles in the dike centre (Fig. 2.24).

The inclined dike segment B2 intrudes the lahar deposits of the country rock, a breccia and dike D (Fig. 2.25). Adjacent to the irregular breccia contact, the trachyte contains angular breccia xenoliths (up to 3cm), the proportion of which decrease from 10% at the contact to 0% four metres from the contact. Xenoliths of lahar deposits (up

to 1cm) occur in the trachyte proximal to the regular lahar contact. Adjacent to dike segment B2, the characteristic yellow to red colour of the lahar deposits is replaced by a grey to red colour. The well developed cooling joints perpendicular to the dike margins extend across the dike segment B2/dike C contact. Baked/chilled margins characterise the breccia and lahar contacts.

Vertical and horizontal apophyses of Group B trachyte, displaying pinch and swell structures, extend into the breccia (Fig. 2.25) with thin, horizontal trachyte veins (up to 5cm thick) emanating from the vertical apophysis. The sharp, sinuous contact between the apophyses and breccia is characterised by chill textures in the trachyte and a well indurated, baked margin (3 - 4cm thick) in the breccia which is often welded onto the trachyte contact surface.

Variation in colour and vesicularity occur across dike segment B2. The non - vesicular trachyte adjacent to the breccia is a light to medium blue colour with a silvery sheen. At the lahar contact the trachyte is green - brown and contains areas of vesicles (up to 5% of the mode) infilled with zeolites. In the horizontal apophyses, the light grey to white trachyte has a vesicularity which constitutes up to 30% of the rock in contrast to the medium grey trachyte in the vertical apophysis which is non - vesicular.

2.4.4 Dike C (Low silica benmoreite)

Dike C (Fig. 2.25) is a moderately indurated, green - brown or light grey, low silica benmoreite with a silvery sheen. Elongate, often sinuous, zeolite - infilled vesicles (up to 2cm) constitute up to 30% of the rock (Fig. 2.26). The sharp, but irregular contact between the inclined dike C and dike segment B2 (Fig. 2.25) is characterised by the absence of baked/chilled margins. The well developed cooling joints in dike C have similar orientations as those in the dike segment B2.

2.4.5 Dike D (Basalt)

The vertical dike D trends 350° west of north and is a well indurated, black, porphyritic basalt (Fig. 2.1, back pocket). Phenocrysts of plagioclase (av 0.3cm), clinopyroxene (av 0.1cm) and olivine (av 0.1cm), constituting respectively 20%, <1%, <1% of the rock, occur in a medium groundmass. Large phenocrysts of plagioclase (up to 1cm) are rare and cooling joints are well developed across the dike width.

A breccia composed of matrix - supported basalt and trachyte clasts (0.3cm - 30cm) outcrops underneath dike D and adjacent to the Group B trachyte apophyses of dike segment B2 (Fig. 2.25). This breccia is predominantly well indurated, but extensive fracturing of many basalt clasts has produced unstable, crumbling areas. Stable areas of breccia tend to contain a higher proportion of matrix and smaller sized

clasts. The yellow, subrounded, moderately indurated trachyte clasts have rims (1 - 2cm wide) of dark orange and comprise up to 10% of the breccia. The well indurated, dark grey to black, angular to subrounded basalt clasts are often surrounded by a thin (<1mm thick), yellow brown, well indurated "skin" and/or a black, well indurated "rind" of glassy material (up to 5cm thick) which breaks conchoidally. Conchoidal fractures are also evident in the black, well indurated matrix, which is composed of angular clasts (up to 0.3cm) and volcanic glass. Adjacent to the trachyte apophyses the breccia matrix is moderately indurated and often crystal rich (up to 30% of the mode).

2.4.6 Dike E (Low silica benmoreite)

The textures and composition of the dike E low silica benmoreite are similar to those described in dike C and will not be discussed further.

The sharp regular contact between dike E and dike D (Fig. 2.1, back pocket) exhibits baked margins in the basalt (up to 2cm wide) and chilled, crenulated flow foliation (up to 5cm wide) in the low silica benmoreite. Baked/chilled margins are also apparent adjacent to the sharp, but irregular breccia contacts (Fig. 2.27). Irregular cooling joints are developed across dike E and these are offset at the dike margins adjacent to the breccia by closely spaced jointing orientated parallel to the contact (Fig. 2.27).

Sinuous apophyses of dike E material extend into the breccia for distances up to 2m.

2.4.7 Dike F (Basalt)

Dike F intrudes the Group B trachyte in Castle Rock (Fig. 2.1, back pocket) with a sharp but sinuous contact. Vertical flow foliation trending parallel to the dike margins is evident across the dike width. The well developed cooling joints (15 - 30cm apart) are sinuous in the dike centre and adjacent to the dike contact the trachyte exhibits a baked margin (1cm wide) and extensively crenulated flow foliation. This aphanitic, dark grey to black, basalt is porphyritic with phenocrysts of plagioclase (up to 2mm), pyroxene (up to 1mm) and olivine (up to 0.5mm), constituting 2%, 4% and 9% of the rock respectively, occurring in a fine groundmass.

2.4.8 Interpretation of field evidence.

The field relationships of dikes B, C, D and E indicate that the basalt dike (D) was intruded first. The disintegration of this dike down into the breccia, the reasonable compositional similarity between the dike and the relatively fragment - free areas of the breccia matrix, and the compositional similarities between the clasts in the lahar deposits and those in the breccia suggest that the basalt dike and the lahar deposits were involved in the

development of the breccia. It is proposed that fuel - coolant reactions, produced by the intrusion of hot basalt magma into wet lahar deposits, brecciated the lahar clasts and matrix. Mixing of this matrix with the rapidly cooled basalt magma produced the glassy breccia matrix, and the reaction rims and "skin" around the trachyte and basalt clasts respectively were the result of hydrothermal activity during these fuel - coolant reactions. Rapid cooling of the magma adjacent to the basalt clasts produced the glassy "rinds".

The blunt termination of dike segment B1 and the occurrence of Group A trachyte in Area B suggests that the intrusion of this trachyte was accompanied by offsetting of the fracture. The trend of the western margin of dike segment B2 and the close relationship between the cooling joints of this segment and dike C implies that these subsequent intrusions utilised the same fracture. The absence of Group B trachyte and low silica benmoreite in dike segment B1 suggests that their intrusion was at deeper levels and this hypothesis is supported by the occurrence of Group B trachyte and low silica benmoreite at lower elevations in Area B (Fig. 1.6).

The xenoliths in the Group B trachyte were derived from the adjacent breccia and lahar deposits during the emplacement of the trachyte. Significant cooling of these units prior to the trachyte intrusion is implied by the angularity of the xenoliths and the presence of the baked/chilled margins at the contacts. The stresses produced by magma velocity changes at the sharp bend in the dike

configuration at this locality may have facilitated the breakup of the well indurated breccia.

The absence of Group A trachyte adjacent to the dike segment B2 suggests not only a close temporal relationship between the two trachytes (Groups A and B), but that the intrusion of the Group B trachyte was accompanied by widening of the fracture.

A close temporal relationship between the Group B trachyte dike segment B2 and the low silica benmoreite dike C is indicated by the absence of chilled/baked margins at the contact and the continuation of the cooling joints through the contact. No inference can be made on the temporal and spatial relationships between the low silica benmoreite dike E and the trachyte dike segment B2. However, given the geochemical similarities between dikes E and C (Chapter 5) it is reasonable to suggest that they were emplaced at about the same time.

Laboratory experiments (McGuire & Pullen, 1989) involving the injection of water (representing magma) from below into solidified gelatin moulds (representing volcanoes) have simulated the formation and propagation of radial dikes. During these experiments, water - filled fractures propagated horizontally from the water column to the surface of the gelatin mould when the finite yield strength of the gelatin was exceeded. After conducting several experiments utilising different mould shapes, McGuire and Pullen (1989) found that a regular conical gelatin mould produced radial fractures resembling those formed during volcanic activity.

The results of these laboratory simulations (McGuire &

Pullen, 1989) suggest that magmatic pressure located around a central volcanic vent in a regular shaped volcanic cone may have initiated the fissure system associated with the feeder dikes of Castle Rock and the Eastern Sheet. As the pressure of the rising magma column exceeded the lithostatic load (tensile strength of the surrounding country rock plus the minimum compressive stress in the horizontal plane), the fracturing conduit walls produced horizontally propagating, vertical fissures, the orientation of which were controlled by the cone morphology (McGuire & Pullen, 1989). A reduction in the lithostatic load at the volcano flanks enabled these propagating fissures to laterally and vertically expand, thus facilitating the emplacement of Castle Rock and the Eastern Sheet. Lavas flows and breccias associated with extruded domes (MacDonald, 1972, Williams, 1932, Price & Coombs, 1975) are not found at Castle Rock and the Eastern Sheet, suggesting that the emplacement of these bodies occurred at shallow depths close to or at the surface.

Displacement attributable to regional movement is not discernable along the lengths of the feeder dikes. This may reflect the difficulty with which the basement stresses, present at the time of volcanism, were generated up through the volcanic deposits (McGuire & Pullen, 1989) and/or the overriding of the regional stress regime by the development of the radial fracture system.

2.5 Synthesis

The close spatial and temporal relationships between the trachytes, high silica benmoreite and low silica benmoreite suggest that they were rapidly erupted from a single magma reservoir during one event. This hypothesis is supported by the relatively continuous cooling joints throughout the trachytes and benmoreites. Each compositional unit (trachytes, high silica benmoreite, low silica benmoreite) was emplaced quickly following one another, however a small proportion of Group B trachyte and high silica benmoreite was intruded after and/or during the emplacement of the low silica benmoreite. The absence of baked/chilled margins and the continuity of the flow foliation throughout the trachyte of Castle Rock, suggests that multiple pulses of trachyte magma injection either did not occur or, if they did, followed each other quickly.

The spatial and temporal relationships between dikes B and C suggest that these were the feeders for the Castle Rock intrusion and this hypothesis is supported by geochemical, petrographical and mineralogical data presented in Chapters 3 and 5. Dike E is also considered a probable feeder for the Castle Rock low silica benmoreite. These feeder dikes were initiated and propagated by a magmatic stress regime, developed around a central conduit in a regular volcanic cone.

The emplacement of Castle Rock and the Eastern Sheet is associated with the later stages of activity at the Lyttelton

Volcano, and occurred at shallow depths or close to the surface on the northern flanks.

3. Petrography and mineralogy

3.1 Introduction

In this chapter the petrography and mineralogy of Castle Rock is presented and described. Each main rock type (trachyte, high silica benmoreite, low silica benmoreite) is described separately and a separate section is allocated to the dike B trachyte. Sample localities are given on Fig 1.6 and sample descriptions are presented in Appendix 1.

Thirty six thin sections were examined: 16 from the trachyte, 3 from the high silica benmoreite, 6 from the low silica benmoreite, 1 from a xenolith within the Castle Rock trachyte, 8 from dike B, 1 from dike A and 1 from dike F. Thin sections from the xenolith, and dikes A and F are omitted from the following discussion. Full petrographic descriptions of all thin sections examined are given in Appendix 2.

The proportions of the mineral phases present in each thin section were determined using the visual estimation charts of Terry and Chilinger (1955). Geochemical criteria, presented in Section 1.6 were used to establish rock nomenclature. Plagioclase anorthite contents, quoted in Appendix 2, were determined using the Michel-Levy method described in Shelley (1985) and are minimum estimates only. The mineral phases have been described as either groundmass or phenocrysts. No distinction has been made between microphenocrysts and phenocrysts. The textural term "train" is used to describe the arrangement of inclusions defining

the core and rim of a phenocryst. The other textural terms used in the petrographic descriptions follow MacKenzie et al (1982).

Fourteen microprobe sections were examined: 5 from the Castle Rock trachyte, 3 from the high silica benmoreite, 3 from the low silica benmoreite, and 3 from the dike segment B2 trachyte. Minerals analysed include plagioclase, alkali feldspar, clinopyroxene, olivine, Fe-Ti oxides and amphibole. To facilitate the interpretation of the core/rim analyses, only selected examples are depicted on the core/rim graphs. Gaps in the compositional ranges of the phenocrysts and groundmasses of a particular mineral are probably due to insufficient microprobe data. Appendix 3 contains the mineral analyses and a description of the microprobe analytical techniques.

The feldspar analyses are plotted on the triangular system $CaAl_2Si_2O_8$ - $NaAlSi_3O_8$ - $KAlSi_3O_8$ and classified according to Deer, Howie and Zussman (1977) Fig. 91. The calcic clinopyroxenes are classified according to the scheme of Morimoto et al (1988) and the analyses are plotted on the pyroxene quadrilateral. A limited number of olivine and amphibole analyses were obtained. The olivine analyses are plotted on the linear Fo - Fa graph and classified according to Deer, Howie and Zussman (1977) Fig. 3 and the classification scheme of Leake (1978) was used to determine the amphibole types. The Fe-Ti oxides are plotted on the triangular system FeO - TiO_2 - Fe_2O_3 and classified according to Deer, Howie and Zussman (1977) Fig. 156.

3.2 Trachyte

3.2.1 Introduction

The trachytes are aphanitic, generally holocrystalline and porphyritic with phenocrysts of alkali feldspar, clinopyroxene, Fe - Ti oxides, apatite and nepheline in trachytic and intergranular groundmasses. The phenocrysts of alkali feldspar, clinopyroxene, Fe - Ti oxides and apatite comprise up to 3% of the mode and occur in the following relative abundances: alkali feldspar > clinopyroxene > Fe - Ti oxides > apatite. The nepheline phenocrysts have been omitted from this list of relative abundances due to their wide variation of occurrence in the mode (0 - 2%). Some of the trachytes contain a small proportion of yellow - brown volcanic glass, most of which is devitrified.

The phenocrysts occur as discrete crystals or as feldspar dominated glomerocrysts comprising the following assemblages:

(1) Alkali feldspar ± clinopyroxene ± Fe - Ti oxides
(2) Clinopyroxene ± Fe - Ti oxides

However, the clinopyroxene - dominated glomerocrysts are rare.

3.2.2 Alkali feldspar phenocrysts

3.2.2.1 Petrography

The euhedral to anhedral alkali feldspar phenocrysts range in size from 0.1mmx0.09mm to 4mmx1mm. Rare large phenocrysts reach lengths of 7mm. The phenocrysts display carlsbad twinning and marginal zoning patterns. Patch zoning is found only in the larger sized phenocrysts and albite (Fig. 3.1) and cross - hatch twinning is rare.

The euhedral to subhedral phenocrysts of the Group A trachyte are anorthoclase (An 3.1 Ab 62.6 - An 4.0 Ab 68.0). These phenocrysts are predominantly inclusion free, although some do contain inclusions of Fe - Ti oxides, clinopyroxene and Fe - Ti oxide pseudomorphs. Glomerocrysts and anhedral phenocrysts are rare.

Three populations of anorthoclase phenocrysts (An 1.1 Ab 62.7 - An 11.6 Ab 73.4) are found in the Group B trachyte. As all three populations occur over this compositional range, the subdivision into populations was based on textural evidence. The first population of phenocrysts have resorbed, sieve - textured cores which often contain inclusions of Fe - Ti oxides, clinopyroxene and Fe - Ti oxide pseudomorphs. These cores are surrounded by subhedral to anhedral clear rims (Fig. 3.2). The resorbed clear cores of the second population of phenocrysts are surrounded by anhedral rims defined by inclusion trains (Fig. 3.3). Many of these rims contain inclusions of clinopyroxene and vermicular Fe - Ti oxides. The main textural difference, between the phenocrysts

of the third population and those of the second population, is the presence of a sieved area (up to 0.18mm wide) located between the resorbed clear cores and the subhedral to anhedral clear rims (Fig. 3.4).

The first and third populations of phenocrysts in the Group B trachyte occur predominantly as consertal textured glomerocrysts (Figs. 3.4 & 3.5) in contrast to the second population of phenocrysts which occur only as discrete crystals (Fig. 3.3). The locality of the sieved areas in the glomerocrysts of the third population of phenocrysts suggests that they developed subsequent to the growth of the glomerocrysts.

The distribution of the anorthoclase phenocryst populations in the Group B trachyte is related to locality. The third and second populations of phenocrysts are absent from the southern end and northern end of Castle Rock respectively. At the southern end of Castle Rock, the second population of phenocrysts is subordinate to the first. This contrasts with the northern end of Castle Rock, where the first population is subordinate to the third.

3.2.2.2 Mineralogy

Fig. 3.6 illustrates the compositional variation of the anorthoclase phenocrysts in the trachyte of Castle Rock. The phenocrysts in the Group A rocks (Ab 68.7 An 4.0 Or 27.3 - Ab 62.7 An 3.1 Or 34.3). have a narrower compositional range than those in Group B (Ab 62.7 An 1.1 Or 36.2 - Ab 73.4 An

11.6 Or 14.9).

Normal and reverse marginal zoning is present in both rock groups. However, the weakly developed marginal zoning in the Group A rocks contrasts with the predominantly strong marginal zoning in the three populations of phenocrysts in the Group B rocks. The anorthoclase phenocryst cores in the Group B trachyte (Fig. 3.7) are rimmed by anorthoclase (Ab 65.1 An 3.0 Or 31.9 - Ab 69.1 An 5.8 Or 25.0) which overlaps in composition with the groundmass (Fig. 3.8). The Ab - rich cores tend to be normally zoned whereas reverse zoning is usually exhibited by the Or -rich cores.

3.2.3 Clinopyroxene phenocrysts

3.2.3.1 Petrography

The euhedral to anhedral clinopyroxene phenocrysts (0.14mmx0.1mm - 1.1mmx0.5mm) display patch, sector and marginal zoning patterns.

The phenocrysts of the Group A trachyte are euhedral to subhedral hedenbergite. These weakly to moderately pleochroic phenocrysts are yellow - green, brown - green, light green or medium green, and sometimes contain inclusions of Fe - Ti oxides.

Two populations of clinopyroxene phenocrysts are present in the Group B trachyte. The first population are non - pleochroic, light green, colourless or pink, subhedral to anhedral diopside phenocrysts and range in size from

0.56mmx0.25mm to 0.9mmx0.56mm. The light green phenocrysts contain inclusions of Fe - Ti oxides, whereas the colourless to pink phenocrysts have resorbed, relatively inclusion - free cores surrounded by green, anhedral, diopsidic rims riddled with inclusions of vermicular Fe - Ti oxides (Fig. 3.9).

The second population of clinopyroxene phenocrysts in the Group B trachyte (0.14mmx0.1mm - 1.5mmx0.4mm) are similar in appearance and composition to the hedenbergite phenocrysts present in the Group A rocks.

3.2.3.2 Mineralogy

Fig. 3.10 illustrates the compositional variation of the clinopyroxene phenocrysts in the Castle Rock trachyte. Hedenbergite (En 14.3 Wo 46.8 Fs 38.8 - En 25.0 Wo 45.8 Fs 29.2). is the only clinopyroxene present in the Group A rocks whereas in the Group B trachyte, both hedenbergite and diopside are found (En 22.9 Wo 45.6 Fs 31.5 - En 38.4 Wo 47.6 Fs 14.0). Fe - rich hedenbergite is absent from the Group B rocks, and the Mg - rich hedenbergite present, compositionally overlaps with the hedenbergite phenocrysts in the Group A rocks.

Marginal zoning is strongly developed in the clinopyroxene phenocrysts of the Groups A and B trachyte (Fig. 3.11). The reverse zoning in the Group A rocks produces hedenbergite rims (En 19.0 Wo 46.6 Fs 34.4 - En 21.0 Wo 45.6 Fs 33.4) which overlap in composition with the groundmass

(Fig. 3.12, A). The Fe - rich diopside rims (av. En 34.9 Wo 47.9 Fs 17.2) surrounding the Mg - rich diopside cores in the Group B trachyte contrast with the Mg - rich hedenbergite (av. En 25.0 Wo 44.7 Fs 30.3) rims of the Fe - rich diopside cores. The reversely zoned hedenbergite phenocrysts in the Group B trachyte have rims of Fe - rich diopside (En 30.3 Wo 45.3 Fs 24.4 - En 26.8 Wo 46.8 Fs 26.5) which overlap in composition with the groundmass (Fig. 3.12, B).

3.2.4 Other phenocryst phases

The nepheline phenocrysts (0.03mmx0.02mm - 0.2mmx0.15mm) in the trachytes are predominantly euhedral although subhedral forms are found. These crystals are rimmed by groundmass clinopyroxene and/or aenigmatite and are altered to clays and zeolites.

The Fe - Ti oxide phenocrysts in the Groups A and B trachyte (0.05mmx0.05mm - 0.28mmx0.18mm) are euhedral to anhedral titanomagnetite (Fig. 3.13 C, D). In the Group B trachyte, the high TiO_2 content in titanomagnetite inclusions within diopside phenocrysts suggests that these inclusions and/or phenocrysts could have been derived from the benmoreites.

The rare, anhedral, pale yellow olivine phenocrysts (0.56mmx0.35mm) are ferrohortonolite (av Fa 84), (Fig. 3.14). Most of these phenocrysts have reaction rims of, (Fig. 3.15) or are pseudomorphed by, Fe - Ti oxides. Some of the olivine phenocrysts are altered to clays.

The colourless, hexagonal apatite phenocrysts (0.02mmx0.02mm - 0.3mmx0.25mm) are euhedral.

3.2.5 Groundmass

The groundmass crystal size in the trachytes increases away from the country rock contacts, and the smaller and larger crystal sizes in the following text pertain to the groundmass adjacent to the country rock and the overlying benmoreite contacts respectively.

The groundmasses of the trachyte rocks are predominantly alkali feldspar with subordinate intergranular clinopyroxene, aenigmatite and euhedral to subhedral titanomagnetite (Fig. 3.13 C, D) (0.01mmx0.01mm - 0.05mmx0.05mm).

The euhedral to subhedral feldspar laths (0.15mmx0.01mm - 0.75mmx0.08mm) form a kinked trachytic texture (Fig. 3.16) which swirls around the phenocrysts. Carlsbad twinning and marginal zoning is displayed in many of the larger sized laths. These groundmass feldspars range in composition from anorthoclase to sanidine (Ab 58.9 An 2.2 Or 38.8 - Ab 70.8 An 13.3 Or 16.0). In each rock group, the compositional ranges of the groundmass and phenocrysts overlap (Fig. 3.8), resulting in a narrower compositional range in the groundmass feldspars of the Group A rocks compared with that present in the groundmass of Group B.

The groundmass clinopyroxenes in the Group A rocks are weakly to moderately pleochroic, yellow - green, light green or medium green hedenbergite (En 15.1 Wo 44.9 Fs 40.1 - En

21.9 Wo 43.9 Fs 34.2), (Fig. 3.12 A). In the groundmass of the Group B rocks, Fe - rich diopside and Mg - rich hedenbergite (En 24.0 Wo 46.1 Fs 29.9 - En 14.7 Wo 44.1 Fs 41.2), (Fig. 3.12 B), are found. Mg - rich diopside and Fe - rich hedenbergite groundmass clinopyroxene is not present in the Group B trachyte. The groundmass clinopyroxenes occur as subhedral to anhedral crystals (0.06mmx0.02mm - 0.3mmx0.09mm), skeletal crystals (0.09mmx0.03mm - 0.6mmx0.09mm) or microlites. Marginal zoning is displayed in the larger sized crystals and microlites and skeletal crystals predominate at the country rock and benmoreite contacts respectively.

The strongly pleochroic, light to dark brown aenigmatite is often opaque and usually rims the groundmass Fe - Ti oxides. It occurs either, as subhedral to anhedral crystals (up to 0.03mm) or as microlites.

3.2.6 Zeolitisation

In the Group B trachyte at the northern end of Castle Rock, up to 15% of the groundmass is replaced by spherulitic zeolites (Fig. 3.17). These zeolitised areas are defined by thin rims of yellow - brown material (Fig. 3.18). In the Group B rocks, zeolites also occur within the xenoliths and the sparse cracks in the anorthoclase phenocrysts associated with the third population.

3.3 High silica benmoreite

3.3.1 Introduction

The high silica benmoreite is aphanitic, holocrystalline and porphyritic with phenocrysts of alkali feldspar, plagioclase, clinopyroxene, apatite and Fe - Ti oxides in sub - trachytic and intergranular groundmasses. A seriate texture is common in the feldspar and clinopyroxene phenocrysts.

Rare olivine xenocrysts (up to 1mm) in the high silica benmoreite are altered to clays and often contain inclusions of euhedral apatite and Fe - Ti oxides.

The phenocrysts comprise up to 10% of the mode and occur in the following relative abundances: plagioclase > clinopyroxene ≥ alkali feldspar ≥ Fe - Ti oxides ≥ apatite.

These phenocrysts occur either as discrete crystals or as plagioclase or clinopyroxene - dominated glomerocrysts in which a consertal texture is sometimes displayed. The two glomerocryst assemblages are :

(1) Plagioclase ± clinopyroxene ± Fe - Ti oxides ± apatite
(2) Clinopyroxene ± Fe - Ti oxides ± olivine ± apatite.

3.3.2 Feldspar phenocrysts

3.3.2.1 Petrography

Textural and compositional evidence suggests that two

populations of feldspar phenocrysts are present in the high silica benmoreite.

The predominantly euhedral to subhedral phenocrysts (0.38mmx0.3mm to 5mmx2mm) of the first population are calcic labradorite to sodic andesine (An 65.7 Ab 33.0 - An 31.0 Ab 62.4). Albite and carlsbad twinning is displayed in these phenocrysts along with delicate patch, convolute, oscillatory and marginal zoning. The cores, which are often resorbed, are surrounded by subhedral to anhedral rims defined by inclusion trains of Fe - Ti oxides and clinopyroxene. Rare inclusions are found within these cores and embayments are common (Fig. 3.19).

The second population of phenocrysts (0.38mmx0.1mm - 3mmx2mm) are carlsbad twinned, subhedral to anhedral oligoclase and anorthoclase (An 23.3 Ab 68.4 - An 8.8 Ab 70.0), displaying patch and marginal zoning patterns. These phenocrysts have resorbed, sieve - texture (Fig. 3.20) cores, often containing inclusions of Fe - Ti oxides and clinopyroxene, surrounded by clear, subhedral to anhedral rims. The textural similarities, between these phenocrysts and the first population of phenocrysts in the Group B trachyte (Section 3.2.2.1), are accompanied by a compositional overlap (Section 3.3.2.2).

3.3.2.2 Mineralogy

Fig. 3.21 (A) illustrates the compositional ranges of the two feldspar phenocryst populations in the high silica

benmoreite. The compositional range of the labradorite to andesine phenocrysts of the first population (Ab 33.0 An 65.7 Or 1.3 - Ab 62.4 An 31.9 Or 5.6) is similar to that in the plagioclase phenocrysts of the low silica benmoreite (Section 3.4.2.2). The oligoclase and anorthoclase phenocrysts of the second population (Ab 68.4 An 23.3 Or 8.4 - Ab 70.0 An 8.8 Or 21.3) compositionally overlap with the feldspar phenocrysts in the first population of the Group B trachyte (Section 3.2.2.2).

Normal marginal zoning is present in both populations of phenocrysts (Fig. 3.22). These plagioclase and anorthoclase cores are surrounded by anorthoclase rims (Ab 73.7 An 12.6 Or 14.1 - Ab 68.3 An 8.2 Or 23.5) similar in composition to the groundmass (Fig. 3.21 B). The smaller variation in composition between the cores and the rims of the second population contrasts with the wider compositional variation between the cores and rims of the first population.

3.3.3 Clinopyroxene phenocrysts

3.3.3.1 Petrography

Two populations of clinopyroxene phenocrysts are present in the high silica benmoreite. Marginal zoning is found in both populations.

The first population are euhedral to subhedral, non - pleochroic, colourless, pink or light green diopside. The colourless to pink phenocrysts (0,75mmx0.75mm - 1.5mmx0.8mm)

have resorbed cores surrounded partially or completely by euhedral to subhedral rims (Fig. 3.23). Inclusions of plagioclase, Fe - Ti oxides and apatite occur in many of these cores. The light green phenocrysts (0.2mmx0/09mm - 0.5mmx0.2mm) contain Fe - Ti oxide inclusions. The diopside phenocrysts have rims of either light green, non - pleochroic diopside or light green, moderately to weakly pleochroic hedenbergite. This variation in rim composition is defined texturally. The diopside rims are riddled with vermicular Fe - Ti oxides and feldspar (Fig. 3.25) in contrast to the hedenbergite rims which contain rare inclusions of Fe - Ti oxides. A large phenocryst of colourless clinopyroxene (4mmx3mm) in sample 13656 contains inclusions of fresh kaersutite and Fe - Ti oxides (Fig. 3.24).

The second population of clinopyroxene phenocrysts (0.18mmx0.09mm - 0.75mmxo.38mm) are subhedral to anhedral, weakly to moderately pleochroic, yellow - green to light green hedenbergite. These phenocrysts have resorbed cores, surrounded by inclusion free subhedral to anhedral rims of either diopside or hedenbergite.

3.3.3.2 Mineralogy

Fig. 3.26, (A) illustrates the compositional variation in the clinopyroxene phenocrysts of the high silica benmoreite. The hedenbergite phenocrysts display a narrower compositional range (En 22.2 Wo 45.4 Fs 32.4 - En 26.3 Wo 46.7 Fs 27.0) compared to that present in the diopsides (En

22.2 Wo 45.4 Fs 32.4 - En 39.0 Wo 46.4 Fs 14.6). The compositional ranges present in these diopside and hedenbergite phenocrysts is similar to that present in the clinopyroxene phenocrysts of the Group B trachyte.

Strong normal and reverse marginal zoning is restricted to the diopside and hedenbergite phenocrysts respectively (Fig. 3.27). The rims of these phenocryst cores fall into two compositional groups: Fe - rich diopside (average En 33.0 Wo 46.0 Fs 20.0), which overlaps in composition with the clinopyroxene groundmass (Fig. 3.26, B), and Mg - rich hedenbergite (average En 25.0 Wo 46.0 Fs 30.0). The Mg - rich and Fe - rich diopside cores are rimmed by an Fe - rich diopside and a Mg - rich hedenbergite respectively whereas the hedenbergite cores have rims of either, a more Mg rich hedenbergite or an Fe - rich diopside.

The complicated growth history of a strongly zoned hedenbergite phenocryst is illustrated on Fig. 3.28. Unfortunately, only the inner core and outer rim of this phenocryst were analysed by microprobe, so the compositional changes occurring in the intermediate zones can not be assessed. The morphological changes illustrated on Fig. 3.28 suggest that four major stages occurred during the growth history of this hedenbergite phenocryst. A resorption stage, represented by the rounded inner core of area 1, was followed by a period of increasing re - equilibration between the phenocryst and the surrounding magma. This re - equilibrium phase, which produced the subhedral area 2 and euhedral area 3 was succeeded by a resorption stage represented by the embayed outer area 4.

3.3.4 Other phenocryst phases

The colourless, euhedral, hexagonal apatite phenocrysts (0.09mmx0.09mm - 0.18mmx0.14mm) in the high silica benmoreite sometimes occur as inclusions in the phenocrysts of clinopyroxene and Fe - Ti oxides.

The predominantly euhedral to subhedral Fe - Ti oxide phenocrysts (0.09mmx0.9mm - 0.4mmx0.5mm) are titanomagnetite (Fig. 3.13 B). Embayments produce anhedral forms, up to 0.35mm, in some of these phenocrysts.

3.3.5 Groundmass

The high silica benmoreite groundmass is predominantly alkali feldspar with subordinate, intergranular clinopyroxene, olivine and Fe - Ti oxides.

The feldspar laths (up to 0.08mmx0.5mm) are euhedral to subhedral anorthoclase (Ab 70.1 An 13.1 Or 16.8 - Ab 61.7 An 3.3 Or 34.9) (Fig. 3.21, B) and display a sub - trachytic texture and carlsbad twinning. The compositions of these groundmass feldspars and the feldspar phenocryst rims overlap.

The clinopyroxenes are light green, non - pleochroic diopside (En 36.6 Wo 44.9 Fs 18.5 - En 33.4 Wo 45.5 Fs 21.1) which overlaps in composition with the Fe - rich diopsidic phenocrysts and rims (Figs. 3.26, A, 3.27). These groundmass clinopyroxenes occur either, as subhedral to anhedral

crystals (up to 0.18mmx0.06mm) or as microlites and the larger crystals sometimes contain Fe -Ti oxide inclusions.

The anhedral, colourless olivine crystals (up to 0.04mmx0.04mm) are hortonolite (av Fa 66) (Fig. 3.14) and have rims which are altered to bowlingite(?).

The Fe - Ti oxides (up to 0.05mmx0.05mm) are euhedral to subhedral titanomagnetite (Fig. 3.13, B).

3.3.6 Amphibole xenocrysts

3.3.6.1 Petrography

Kaersutite xenocrysts in the high silica benmoreite (Figs. 3.24, 3.25) are mantled by reaction rims of Fe - Ti oxides. Holes within many of the crystal centres are infilled with apatite, vermicular feldspar, clinopyroxene and Fe - Ti oxides. These xenocrysts occur predominantly as discrete crystals, but some are attached to the edges of clinopyroxene phenocrysts. A few of the kaersutite xenocrysts are pseudomorphed by Fe - Ti oxides.

The difficulty in distinguishing between phenocrysts which have experienced magma mixing and xenocrysts must be born in mind when determining the origin of disequilibrium textures. The kaersutite crystals in the rocks of the study area have been categorised as xenocrysts due to the presence of Fe - Ti oxide reaction rims, which are considered to be the breakdown products of high pressure amphiboles in a low pressure environment (Stewart, 1975) (Chapter 6).

3.3.6.2 Mineralogy

Only a limited number of amphibole analyses were obtained, one of which is presented in Appendix 3. According to the scheme of Leake (1978) these calcic amphiboles [$(Ca + Na)_B > 1.34$, $Na_B < 0.67$, $Ca_B > 1.34$] are classified as kaersutites [Ti > 0.5, Si > 6.5 and $Mg/(Mg+Fe^{2+}) = 0.6$]. An $Al^{VI} > 1$ suggests alumino - kaersutite.

3.4 Low silica benmoreite

3.4.1 Introduction

The low silica benmoreite is aphanitic, holocrystalline and porphyritic with plagioclase, clinopyroxene, olivine, apatite and Fe - Ti oxide phenocrysts in sub - trachytic and intergranular groundmasses. A seriate texture is common in the plagioclase and clinopyroxene phenocrysts. The phenocrysts comprise up to 15% of the mode and occur in the following relative abundances: plagioclase > clinopyroxene > olivine ≥ Fe - Ti oxide > apatite.

Rare subhedral to anhedral alkali feldspar xenocrysts (av 1mmx0.5mm) in the low silica benmoreite display sieved cores (Fig. 3.29), often containing inclusions of Fe - Ti oxides and clinopyroxene, surrounded by clear anhedral rims. Texturally these xenocrysts resemble the anorthoclase phenocrysts of the first and second populations in the Group

B trachyte and high silica benmoreite respectively.

Kaersutite xenocrysts are sporadically distributed throughout the low silica benmoreite and are either surrounded by reaction rims of (Fig. 3.30), or pseudomorphed by, Fe - Ti oxides. Holes in the centres of these resorbed xenocrysts are infilled with Fe - Ti oxides, clinopyroxene and vermicular feldspar.

The phenocrysts occur either as discrete crystals or as plagioclase (Fig. 3.31) or clinopyroxene - dominated glomerocrysts in which a consertal texture is often displayed (Figs. 3.31, 3.32). Rare olivine - dominated glomerocrysts are found. These three glomerocryst assemblages are :

(1) Plagioclase ± clinopyroxene ± olivine ± Fe - Ti oxides ± apatite
(2) Clinopyroxene ± Fe - Ti oxides ± apatite ± olivine
(3) Olivine ± Fe - Ti oxides ± apatite

3.4.2 Plagioclase phenocrysts

3.4.2.1 Petrography

The plagioclase phenocrysts (0.1mmx0.1mm - 4mmx2mm) are euhedral to subhedral andesine to labradorite (An 55.0 Ab 42.9 - An 29.4 Ab 64.8). Rare large phenocrysts reach 6mm in length. Carlsbad and albite twinning is present in these phenocrysts along with patch, convolute, oscillatory and marginal zoning patterns (Fig. 3.31). The euhedral to

subhedral, sometimes resorbed phenocryst cores are surrounded by subhedral to anhedral anorthoclase rims. Rare inclusions of olivine, Fe - Ti oxides and clinopyroxene occur in the phenocryst cores. Some of the phenocrysts are skeletal with elongate holes in the centre filled with groundmass and fine, sparse cracks in plagioclase phenocrysts of samples adjacent to the Group B trachyte contact, are infilled with zeolites.

3.4.2.2 Mineralogy

The compositional variation of the plagioclase phenocrysts in the low silica benmoreite (Fig. 3.33, A) is similar to that in the first population of phenocrysts in the high silica benmoreite. These labradorite to andesine cores (Ab 42.9 An 55.0 Or 2.1 - Ab 64.8 An 29.4 Or 5.7) are rimmed by anorthoclase (Ab 61.1 An 5.4 Or 33.5 - Ab 66.7 An 13.1 Or 20.2) (Fig. 3.34) which compositionally overlaps with the groundmass feldspar (Fig. 3.33, B).

3.4.3 Clinopyroxene phenocrysts

3.4.3.1 Petrography

The euhedral to subhedral clinopyroxene phenocrysts (0.25mmx0.18mm - 2.5mmx1.5mm) are non - pleochroic, colourless, pink or light green diopside. Anhedral crystals are rare. Delicate patch, convolute, sector and marginal

zoning patterns are present in these phenocrysts.

Many of these clinopyroxene phenocrysts contain inclusions of vermicular to subhedral plagioclase, euhedral apatite and Fe - Ti oxides (Fig. 3.35).

3.4.3.2 Mineralogy

The compositional variation in the diopside phenocrysts (En 34.2 Wo 46.9 Fs 18.9 - En 39.8 Wo 47.0 Fs 13.2) (Fig. 3.36, A) is similar to that in the Mg - rich diopside phenocrysts of the first populations in the Group B trachyte and the high silica benmoreite. The cores are surrounded by diopside rims (En 42.4 Wo 45.7 Fs 11.9 to En 36.5 Wo 44.8 Fs 18.8) which overlap in composition with the groundmass (Fig. 3.36, B).

3.4.4 Other phenocryst phases

The euhedral, apatite phenocrysts (0.05mmx0.05mm - 0.38mmx0.18mm) in the low silica benmoreite are colourless and often occur as inclusions in the clinopyroxene, plagioclase and Fe - Ti oxide phenocrysts.

The subhedral to anhedral olivine phenocrysts (0.1mmx0.09mm - 1mmx0.55mm) are hyalosiderite. These normally zoned phenocryst cores (Fa 38.5 - 46.9) are surrounded by rims ranging in composition from Fa 46.5 - 47.6 (Fig. 3.14). The rims and cracks of these phenocrysts are altered to

bowlingite(?) (Fig. 3.38).

The Fe - Ti oxide phenocrysts are euhedral to subhedral (0.1mmx0.1mm - 0.6mmx0.6mm) or anhedral (up to 0.9mm) titanomagnetite (Fig. 3.13, A). Some of these phenocrysts contain exsolution lamellae of ilmenite (Fig. 3.39). Interstitial, granular chalcopyrite (tr), containing anhedral pyrrhotite was found in sample 13624 (Fig. 3.40).

3.4.5 Groundmass

The groundmass of the low silica benmoreite is predominantly alkali feldspar with subordinate, intergranular clinopyroxene, olivine and Fe - Ti oxides.

The euhedral to subhedral, carlsbad and albite twinned feldspar laths (up to 0.38mmx0.03mm) are anorthoclase and sanidine (Ab 66.2 An 10.0 Or 23.8 - Ab 58.5 An 3.9 Or 37.6) (Fig. 3.33, B) and display a sub - trachytic texture that swirls around the phenocrysts. These groundmass feldspars overlap in composition with the rims of the strong marginally zoned phenocrysts.

The light green, non - pleochroic, subhedral to anhedral clinopyroxenes (up to 0.14mmx0.04mm) are diopside (En 33.1 Wo 46.6 Fs 20.3 - En 38.1 Wo 45.8 Fs 16) (Fig. 3.36, B), and together with the anhedral olivine (up to 0.09mmx0.05mm) and euhedral to subhedral Fe - Ti oxides (up to 0.03mmx0.03mm) form an intergranular texture. The compositions of the groundmass and phenocryst diopsides overlap, but a narrower compositional range is displayed by the groundmass. The

hortonolite (Fa 54 - 58) olivine crystals are more Fe - rich than the olivine phenocryst cores and rims, but are less Fe - rich than the groundmass olivine in the high silica benmoreite (Fig. 3.14). The Fe - Ti oxides are titanomagnetite (Fig. 3.13, A).

3.5 Trachyte dike B

3.5.1 Introduction

The trachyte of dike segment B1 is similar in texture and composition to the Group A trachyte described in Section 3.2 and will not be discussed further.

The Group B trachyte of dike segment B2 is aphanitic, holocrystalline and porphyritic, with phenocrysts of alkali feldspar, plagioclase, clinopyroxene and Fe - Ti oxides in trachytic and intergranular groundmasses. A seriate texture is common in the feldspar and clinopyroxene phenocrysts. The phenocrysts comprise up to 3% of the mode and occur in the following abundances: alkali feldspar > clinopyroxene ≥ plagioclase ≥ Fe - Ti oxides.

Iddingsitised cracks and rims occur in the rare, anhedral, olivine xenocrysts and the compositional overlap of these hyalosideritic olivines (Fa 42) with the olivine phenocrysts in the low silica benmoreite (Fig. 3.14) suggests that the former were derived from the latter.

Resorbed kaersutite xenocrysts (up to 0.8mm) are sporadically distributed throughout the dike segment B2

trachyte and occur predominantly as discrete grains, although some are attached to the clinopyroxene phenocrysts. The reaction rims of Fe - Ti oxides surrounding these xenocrysts are thinner than those surrounding the kaersutite xenocrysts in the benmoreites.

The phenocrysts in the dike segment B2 trachyte occur predominantly as discrete grains. Rare plagioclase or clinopyroxene - dominated glomerocrysts display a consertal texture.

3.5.2 Feldspar phenocrysts

3.5.2.1 Petrography

Three populations of feldspar phenocrysts are present in the dike segment B2 trachyte. The first population comprises subhedral to anhedral plagioclase (0.38x0.35mm - 3mmx2mm) which ranges in composition from labradorite to andesine (An 1.9 Ab 59.9 - An 53.2 Ab 45.2). These phenocrysts display carlsbad and albite twinning along with patch, convolute, oscillatory and marginal zoning patterns. Many of these phenocrysts are skeletal and the elongate holes in the crystal centres are infilled with groundmass and spherulitic zeolites. Inclusions of Fe - Ti oxides, hyalosideritic olivine (av Fa 38) and clinopyroxene are rare.

The second population are subhedral to anhedral anorthoclase (0.35mmx0.25mm - 3.5mmx1mm) which ranges in composition from An 10.9 - An 1.9. Carlsbad and albite

twinning is present in these phenocrysts along with patch and marginal zoning patterns. Cross - hatch twinning is rare. The clear, resorbed cores of these phenocrysts are surrounded by subhedral to anhedral rims which are often defined by inclusion trains of Fe - Ti oxides and pyroxene. Exsolution lamellae and blebs are present in most of these phenocrysts.

The third population comprises phenocrysts of plagioclase (up to 2mmx1.5mm) and anorthoclase (Fig. 3.41), (up to 3mmx1mm), with sieved, resorbed cores surrounded by clear, subhedral to anhedral rims defined by inclusion trains of Fe - Ti oxides. These phenocryst cores often contain inclusions of clinopyroxene and Fe - Ti oxides.

3.5.2.2 Mineralogy

The geochemical variation of the three populations of feldspar phenocrysts in the dike segment B2 trachyte is illustrated on Fig. 3.42. The labradorite and andesine phenocrysts of the first population (Ab 55.0 An 43.4 Or 1.6 - Ab 45.2 An 53.2 Or 1.6) overlap in composition with the plagioclase phenocrysts in the benmoreites (Figs. 3.21, A & 3.33, A). The compositional similarity between the olivine inclusions in the plagioclase phenocrysts of the dike segment B2 trachyte and the olivine phenocrysts of the low silica benmoreite (Fig. 3.14) suggests that the plagioclases in the trachyte were derived from the benmoreite. The anorthoclase phenocrysts of the second population (Ab 75.0 An 10.9 Or 14.1 - Ab 59.9 An 1.9 Or 38.2) are texturally and compositionally

similar to the second population of anorthoclase phenocrysts in the Group B trachyte (Fig. 3.3, 3.6, B). The sieved, oligoclase and anorthoclase phenocrysts of the third population (Ab 70.0 An 24.5 Or 5.5 - Ab 69.5 An 4.0 Or 26.4) are not only texturally and compositionally similar to the second population of feldspar phenocrysts in the high silica benmoreite (Fig. 3.20 & 3.21, A), but they also overlap in composition with, and have similar textures to the first population of feldspar phenocrysts in the Group B trachyte (Fig. 3.2, 3.6, B).

Marginal zoning is present in all three populations but it is more strongly displayed in the first and third populations (Fig. 3.43). The phenocryst cores of the first population are surrounded by andesine to anorthoclase rims ((Ab 53.1 An 44.7 Or 2.2 - Ab 74.1 An 7.4 Or 18.5). The third population of phenocrysts displays normal and reverse marginal zoning where the oligoclase cores are rimmed by anorthoclase (av Ab 68.6 An 7.6 Or 23.8) in contrast to the anorthoclase cores which are rimmed by oligoclase (av Ab 72.6 An 21.7 Or 5.7). The anorthoclase rims surrounding the phenocryst cores are similar in composition to the groundmass feldspar (Fig. 3.42, B).

3.5.3 Clinopyroxene phenocrysts

3.5.3.1 Petrography

The clinopyroxene phenocrysts are predominantly euhedral

to subhedral, although anhedral forms are found. The zoning patterns displayed are patch and marginal. Two populations of clinopyroxene phenocrysts are present.

The first population is non - pleochroic diopside (En 15.1 - En 28.2). These phenocrysts (0.13mmx0.12mm - 0.56mmx0.14mm) are predominantly light green and marginally zoned with rims defined by inclusion trains of Fe - Ti oxides. The rare, colourless or pink diopside phenocrysts (0.14mmx0.06mm - 2mmx1mm) are usually deeply embayed (Fig. 3.44) but the smaller sized phenocrysts have resorbed cores surrounded by anhedral, light green diopsidic rims which are riddled with vermicular Fe - Ti oxides.

The second population of phenocrysts (0.18mmx0.14mm - 0.38mmx0.25mm) comprises weakly to moderately pleochroic, yellow - green, light green or medium green hedenbergite.

3.5.3.2 Mineralogy

Fig. 3.45 illustrates the compositional variation of the clinopyroxene phenocrysts in the Group B trachyte of dike segment B2. As in the Group B trachyte and the high silica benmoreite of Castle Rock, both hedenbergite (En 24.7 Wo 45.5 Fs 29.8 - En 40.1 Wo 46.4 Fs 13.4) and diopside (En 15.1 Wo 46.5 Fs 38.4 - En 28.2 Wo 44.7 Fs 27.1) clinopyroxenes are present. The compositional similarities between the diopside phenocrysts in the Group B trachyte, high silica benmoreite and dike segment B2 trachyte are accompanied by textural similarities (Section 3.5.3.1).

3.5.4 Fe - Ti oxides

The predominantly euhedral to subhedral Fe - Ti oxide phenocrysts (0.04mmx0.04mm - 0.2mmx0.2mm) are titanomagnetite (Fig. 3.13, E). High TiO_2 and Al_2O_3 contents (24.51wt% 5.76wt% respectively) in some of the phenocrysts and inclusions, suggest that they were derived from the benmoreites.

3.5.5 Groundmass

The segment dike B2 trachyte groundmass is predominantly sub - trachytic and comprises alkali feldspar laths (up to 0.4mmx0.02mm) with subordinate, intergranular clinopyroxene microlites and euhedral to subhedral Fe - Ti oxides (up to 0.02mmx0.02mm). The feldspar laths (Ab 74.5 An 12.2 Or 13.4 - Ab 63.2 An 1.5 Or 35.3) are similar in composition to the anorthoclase rims surrounding the phenocryst cores.

3.5.6 Vesicles

Areas of dike segment B2 contain circular to elongate (often sinuous) vesicles (0.09mm - 3mm) which are infilled with irregularly, intergrown chabazite crystals, the size of which increases towards the vesicle centre.

3.6 Xenoliths

The sparsely distributed (<1% of the mode) subrounded to subangular xenoliths in the high silica benmoreite and the trachytes reach lengths of 8cm. Most of these xenoliths are plagioclase rich, although rare olivine - dominated xenoliths are found in the Group B trachyte. The three xenolith mineral assemblages are :

(1) Plagioclase ± clinopyroxene ± Fe - Ti oxides ± apatite
(2) Olivine (iddingsitised) ± Fe - Ti oxides ± apatite
(3) Highly altered xenoliths in which the determination of the crystal phases present is difficult. Plagioclase can be identified in the less altered xenoliths.

The distribution of these xenolith types is related to locality. Types (1) and (2) occur only in the Group B trachyte and the high silica benmoreite. The occurrence of type (3) is restricted to the Group A trachyte.

Type (1) xenoliths petrographically and compositionally resemble the benmoreites. The labradorite to andesine cores (Ab 54.5 An 42.4 Or 3.1 - Ab 46.6 An 51.1 Or 2.3) of the plagioclases are rimmed by a sodic feldspar. The non - pleochroic, sometimes marginally zoned diopside crystals (En 34.0 Wo 46.2 Fs 19.8 - En 39.0 Wo 46.4 Fs 14.6) are light green, colourless or pink, with the light green crystals often containing inclusions of Fe - Ti oxides.

The textural and compositional similarities between the plagioclase and the diopside crystals in the type (1)

xenoliths and the phenocrysts in the low silica benmoreite suggest that the xenoliths were derived from the benmoreite. The absence of quenched feldspar groundmass within and around the undeformed xenoliths indicates that the xenoliths were partly solidified, but not significantly cooled, when they were incorporated into the magma. The compositional similarity between the sodic feldspar rimming the xenoliths and the groundmass feldspar, suggests that the inclusion of the xenoliths occurred prior to or during the crystallisation of the groundmass. A brief time interval between the inclusion of the xenoliths and the onset of groundmass crystallisation is indicated by the sharp edges of the xenoliths.

The highly altered type (3) xenoliths probably represent country rock incorporated during magma emplacement. The quenched feldspar laths in the trachytic groundmasses surrounding and invading these xenoliths and the absence of deformation structures in the xenoliths indicates that the xenoliths were cold when included in the trachyte.

Three types of subangular to angular xenoliths (0.75mmx0.38mm - 2cmx0.6cm) occur in the Group B trachyte of dike segment B2 adjacent to the breccia contact :

(1) Basalt: (Fe - Ti oxide rich, cryptocrystalline.)
(2) Basalt: (porphyritic with phenocrysts (tr) of plagioclase and Fe - Ti oxides in an oxide - rich groundmass containing feldspar.)
(3) Trachyte: (porphyritic with phenocrysts (tr) of alkali feldspar and clinopyroxene in an oxide - rich, trachytic and

intergranular groundmass containing alkali feldspar laths and clinopyroxene.) These xenoliths resemble trachyte from the sill which occurs as part of the country rock (Fig. 2.1, back pocket).

The petrographical similarity of these xenoliths to the breccia suggest that the former were derived from the latter during the trachyte intrusion.

3.7 Interpretation of petrography and mineralogy

With fractionation, the clinopyroxene phenocrysts in the rocks of the study area exhibit an increase in Fe accompanied by, initially a decrease in Ca followed by an increase in Ca (Fig. 3.46). These trends are similar to those found in high calcic clinopyroxene phenocrysts of other alkaline suites (Fodor et al, 1975, Gibb, 1973). The TiO_2 and Al_2O_3 contents of the clinopyroxene phenocrysts decrease with increasing fractionation and range from 2.07 wt% - 0.79 wt% and 5.27 wt% - 1.54 wt% respectively in the low silica benmoreite to 0.38 wt% - 0.26 wt% and 1.13 wt% - 0.84 wt% respectively in the Group A trachyte. In the high silica benmoreite and Group B trachyte, the clinopyroxene phenocrysts have TiO_2 and Al_2O_3 contents (0.20 wt% - 2.05 wt% and 0.67 wt% - 6.24 wt% respectively) which encompass the entire range of values above, suggesting that some of them could have been derived from the low silica benmoreite and Group A trachyte. The increases in clinopyroxene Fe - content are accompanied by

increases in the Fe - content of the olivine phenocrysts.

The labradorite to anorthoclase trend exhibited by the feldspar phenocrysts in the benmoreites and trachytes (Fig. 3.47) is similar to that found in other alkaline suites (Le Roex & Erlank 1982).

Sieve textures, resorption features and reverse zoning are restricted to the feldspar and clinopyroxene phenocrysts in the high silica benmoreite and Group B trachyte. Such disequilibrium features have been proposed by various authors as evidence of magma mixing (Dungan & Rhodes, 1978, Sakuyama, 1978, Lofgren & Norris, 1981, Barton et al, 1982, Giannetti & Luhr, 1983, Morrice & Gill, 1986, Nixon, 1988).

Sieve textures in feldspar phenocrysts can also be produced by rapid crystal growth (Lofgren, 1974). However, the irregular inner sieved boundaries of the third population of phenocrysts in the Group B trachyte and the cross - cutting relationships between these sieved areas and the crystal twins suggests that resorption, not rapid crystal growth, was the mechanism responsible for the sieved textures in the feldspars (Dungan & Rhodes, 1978, Morrice & Gill, 1986, Sakuyama, 1978). The cross - cutting relationships between the sieved areas and the crystal boundaries also suggests that the resorption event producing the sieved areas followed phenocryst and glomerocryst growth.

Mixing experiments conducted by Tsuchiyama (1985) and Lofgren and Norris (1981) suggest that sieve textures and resorbed cores in plagioclase crystals could be produced by mixing mafic and felsic magmas. During these mixing experiments, the sodic plagioclase in the superheated felsic

magma dissolved, became rounded and developed sieve textures but the calcic plagioclase crystals in the supercooled mafic magma were little affected. In the high silica benmoreite and Group B trachyte, the presence of resorbed sieved cores in the oligoclase and anorthoclase phenocrysts and the absence of sieve textures in the resorbed cores of the andesine and labradorite phenocrysts resemble the textures produced in the mixing experiments, suggesting that mixing of a mafic and felsic magma could have occurred. The mixing of the high temperature diopsides and the low temperature hedenbergites with low temperature and high temperature magmas respectively could also produce the resorbed cores and rims found in the clinopyroxene phenocrysts of the high silica benmoreite and Group B trachyte.

The presence of reverse zoning in the hedenbergite and anorthoclase phenocrysts within the high silica benmoreite and Group B trachyte suggests that mechanisms other than crystal fractionation were operating in the magma chamber. These reverse zoning patterns are most simply explained by changes in temperature and chemical composition as the result of magma mixing (Sakuyama, 1978, Giannetti & Luhr, 1983).

The Group A trachyte and the low silica benmoreite contain one population of feldspar and clinopyroxene phenocryst, anorthoclase/hedenbergite and plagioclase/diopside respectively. The presence of these simple mineral assemblages and the absence of disequilibrium textures suggests that these rocks have evolved by crystal fractionation and have not experienced magma mixing processes. The high silica benmoreite and Group B trachyte

contain multiple populations of feldspar and clinopyroxene phenocrysts trending in composition from plagioclase to anorthoclase and diopside to hedenbergite. These diopside/hedenbergite and plagioclase/ anorthoclase phenocrysts could not have been equilibrium with the same melt composition. The evidence of disequilibrium textures in the feldspar and clinopyroxene phenocrysts of the high silica benmoreite and Group B trachyte, and the compositional similarities between these phenocrysts and the feldspar and clinopyroxene phenocrysts in the low silica benmoreite and Group A trachyte suggests that some of the phenocrysts in the former could have been derived from the latter by magma mixing processes.

4. Microfabric

4.1 Introduction

The microfabrics discernible in the trachytes (Groups A and B), and benmoreites (high silica and low silica) were considered to be synonymous with microfabrics studied in radial dikes of the Lyttelton Volcano (Shelley, 1985). This suggested that the techniques used by Shelley (1985) to establish magma flow directions could be applied to the rocks in the study area.

Shelley (1985) determined paleoflow directions in the radial dikes by measuring the orientations of groundmass feldspar laths. Orientated dike samples were collected with their positions noted relative to the dike wall. Using standard U - stage methods, the poles to (010) planes of feldspar groundmass laths, in thin sections cut parallel to the dike walls, were measured and plotted on the lower hemisphere of an equal area net. These plots produced concentrations of poles to (010) in two diametrically opposite regions of the net (Fig. 4.1) which were considered to represent girdles of poles to the (010) planes. Limitations imposed by the U - stage tilt meant that poles lying at angles greater than 50° from the horizontal could not be measured and consequently a "blind spot" in the centre of the plots was produced (Fig. 4.1). The axes to these girdles of poles to the (010) planes represented the flow lines along which a unique flow direction could not be determined. The second step of Shelley's study involved

plotting poles to (010) planes of groundmass feldspar laths in thin sections cut perpendicular to the dike walls and parallel to the girdle axes of the poles to the (010) planes of the groundmass feldspars in the thin sections cut parallel to the dike walls. These plots established that the laths were oblique to the dike walls and that the acute angle between them and the walls always faced away from the magma source. Shelley (1985) concluded from his study that the axes of girdles of poles to the (010) planes determined flow lines and that the obliquity of these girdles to the dike walls determined flow directions.

A study of the groundmass crenulation prevalent throughout many of the dikes of the Lyttelton Volcano led Shelley (1985) to suggest that it could also be used as a means of determining magma flow directions. Shelley (1985) concluded that the acute angle between the shear planes of these crenulations and the dike walls always faced the proposed magma source. A diagrammatic representation of the relationship between the groundmass laths, groundmass crenulation and magma flow direction as established by Shelley (1985) is illustrated in Fig. 4.2.

The obliquity of the groundmass feldspar laths established by Shelley (1985) is similar to that produced by models simulating the behaviour of rigid particles in flowing magmas. Modelling on the assumptions that plane strain deformation occurs during Newtonian flow and that space is sufficient between adjacent particles to allow particle movement during magma flow, Blanchard et al (1979) predicted that velocity gradients during magma flow could rotate

particles to an oblique orientation relative to the dike walls.

Following the technique of Shelley (1985), an analysis of the flow foliation in the dike segment B1 trachyte, the trachyte of Castle Rock and the Eastern Sheet and the benmoreites of Castle Rock was undertaken. U-stage measurements of elongate groundmass feldspars were assessed to determine magma flow lines. Local magma flow directions were established for some of these flow lines by examining the relationship of the oblique groundmass fabrics to the dike margins and the upper and lower contacts of the trachytes and benmoreites.

The aim of the microfabric measurements was to determine the behaviour of the igneous bodies during emplacement. It was thought that by collecting flow data from a number of localities across the intrusions and supplementing it with an examination of the groundmass textures, that magma flow dynamics could be assessed.

4.2 Methodology of flow azimuth determination

4.2.1 Flow lines

Orientated samples from Castle Rock, the Eastern Sheet and dike segment B1 were collected. The sample localities were chosen so as to represent the vertical and horizontal flow foliation variation. The procedure used for analysis of the platy feldspars followed that recommended by Turner and

Weiss (1963) for sheet silicates, and involved cutting several thin sections from each sample. The flow foliation is the most readily identified fabric in the rocks of the study area and the first thin section from each sample was cut parallel to this. Platy groundmass feldspars were selected for the fabric measurements as they constitute most of the mode and have a high susceptibility to alignment during flow. Their platy shapes are also conducive to reliable orientation measurements.

Standard universal stage methods (Shelley, 1985) were used to measure the perpendiculars to (010) in the feldspars of each thin section. In order to remove bias during fabric measurements, the locality of the points at which these measurements would be taken, was established prior to the study and adhered to throughout. One hundred measurements were taken on each thin section.

To establish a flow line (if one existed), the poles to (010) for each thin section cut parallel to the flow foliation were plotted on the lower hemisphere of an equal area net. Flow lines were considered to be parallel to the axes of the girdles of poles to (010). A second thin section was cut parallel to some of these proposed flow lines to test the consistency of the results obtained from the foliation parallel thin sections. The examination of five of these thin sections produced flow lines parallel to the initial girdle axes and established that accurate determinations of the flow lines could be made from a single thin section cut parallel to the flow foliation. The flow lines for the remaining

twenty five orientated samples were established from these girdle axes.

4.2.2 Flow directions

To determine the local flow directions two methods, Shelley (1985) and Vernon (1987), were applied to those thin sections cut parallel to the girdle axes and cut from samples adjacent to the contacts.

Shelley (1985) proposed that the obliquity of groundmass fabrics to dike walls reflects a unique magma flow direction (Fig. 4.2). The obliquity of the groundmass fabric in dike segment B1 and Castle Rock (trachyte, benmoreite) samples to the dike margins and trachyte and benmoreite upper and lower contacts respectively enabled the local flow directions to be established (Fig. 4.3). Given the assumption that the oblique groundmass fabric in dikes is produced by velocity gradients (Shelley, 1985), the analogy between this obliquity and that in the Castle Rock samples assessed was considered valid, as similar velocity gradients would occur in the moving magma at the upper and lower contacts.

Magma flow directions were also established by assessing the arrangement of the groundmass feldspar laths adjacent to the phenocrysts following the technique of Vernon (1987). The sense of simple shear was determined by the relative rotation of the phenocrysts. The assumption that a stronger resistance to flow occurs at the base and top of a lava flow (Vernon, 1987) and at dike margins (Shelley, 1985) effectively

eliminated one of the shear directions and enabled the local flow direction to be established.

4.3 Results of flow analysis

4.3.1 Groundmass feldspar lath orientations

The groundmass feldspar lath measurements produced concentrations of poles to (010) planes in two diametrically opposite regions of the equal area net. These concentrations were synonymous with those produced by Shelley (1985) and were considered to represent girdles of poles to the (010) planes. A preferred orientation was considered to exist if greater than 40% of the poles fell within two 50° arcs extended from the centre of the nets and encompassing both concentrations. The degree of preferred orientation was measured against the percentage of poles which plotted within these arcs:

(a) Strong preferred orientation: >60%
(b) Moderate preferred orientation: 50% - 60% (inclusive)
(c) Weak preferred orientation: 40% - 49% (inclusive)
(d) No preferred orientation: <40%

Preferred orientations of feldspar groundmass laths were found in 97% of the samples assessed (Figs. 4.4 - 4.33). 73% and 18% of the sample population exhibited a strong preferred orientation and a moderate preferred orientation

respectively. The low silica benmoreite sample 13620, collected 1.20m above the lower contact, produced a weak preferred orientation (Fig. 4.34) in contrast to sample 13644 (low silica benmoreite), collected at the lower contact, which produced no preferred orientation at all (Fig. 4.35). The results of the feldspar measurements for these two samples contrasts with the strong preferred orientation within the low silica benmoreite samples (Figs. 4.31 - 4.33) collected approximately 30m above the lower contact. The disintegration of the strong preferred orientation of the groundmass feldspars in the low silica benmoreite adjacent to the lower contact may reflect the difficulty with which it moved over the Group B trachyte. Alternatively, groundmass growth at the base of the low silica benmoreite subsequent to emplacement could have produced the randomly orientated groundmass feldspar laths. The trachyte sample 13643, collected 8m below the upper contact, displays a weak preferred orientation (Fig. 4.35). The shallow dipping flow foliation (19°) at this locality contrasts with the steeply dipping flow foliation adjacent to the trachyte samples from the middle sections of Castle Rock (13682, 13680) and the Eastern Sheet (13697, 13700) which exhibit strong preferred orientations (Figs. 4.12, 4.14, 4.26, 4.27), and it is suggested that distal from the trachyte upper and lower contacts, a strong preferred orientation may be more difficult to achieve in shallow dipping flows.

The strong preferred orientation exhibited by the groundmass feldspar in the trachytes (Group A and B), high

silica benmoreite and low silica benmoreite is mimicked by the feldspar phenocrysts.

4.3.2 Groundmass obliquity

The determination of flow directions following Shelley's technique requires that the groundmass fabric is oblique to the dike wall. The obliquity of the groundmass laths to the eastern margin of dike segment B1 (Fig. 4.8b) is apparent, but an inspection of Fig. 4.7b suggests that the groundmass feldspars have an immeasurable obliquity to the western dike margin. More U-stage measurements of groundmass feldspars in sample 13712 may have resolved this problem, however this was felt unnecessary, as a flow direction could be determined by an assessment of the groundmass crenulation.

The obliquity of the groundmass feldspars to the lower and upper contact of the Castle Rock trachyte is apparent in Fig. 4.5b and Fig. 4.4b respectively. The obliquity of the groundmass feldspar in the high silica benmoreite sample, 13656, and the low silica benmoreite sample, 13622, were measured against the orientation of the upper contact and the columnar joint ends respectively. As the evidence presented in Section 4.4.3 suggests that the lower surface of the low silica benmoreite was essentially flat, the inclination of the columnar joint ends at the top of Castle Rock were considered to represent the lava flow top.

Vernon's (1987) method for determining flow directions produced results consistent with those determined by the Shelley method.

4.4 Flow dynamics

4.4.1 Introduction

The flow data determined from the feldspar groundmass measurements were assessed in terms of the central, elongate magma source(s) implied by the configurations of Castle Rock and the Eastern Sheet.

Fig. 4.37 illustrates the flow measurements established for the trachytes (Groups A and B) and benmoreites (low silica and high silica).

4.4.2 Trachyte

The trachytes of dike segment B1, Castle Rock and the Eastern Sheet exhibit variable flow data suggesting complex magma flow behaviour during emplacement.

The contrast between the flow data plunges of samples 13628, 13692 and 13704 into the magma source and those of samples 13695 and 13696 away from the Eastern Sheet is consistent with the orientation of the contact plane (Fig. 2.1, back pocket) suggesting an influence of the contact configuration on magma flow behaviour. The upper contact configuration of the Eastern Sheet and the flow data from samples adjacent to it suggest a magma source to the west.

The flow data for trachyte samples distal from the country rock contact (13619, 13686, 13676, 13677) and adjacent to the benmoreite contacts (13621, 13638, 13647) plunges away from the magma source.

At localities in the upper sections of the trachyte, where extreme variability is displayed by the flow data from samples adjacent to each other (13682, 13683 and 13697, 13700), the flow foliation is steeply dipping.

The plunges of the dike segment B1 samples (14038, 13712) illustrate the variability of magma flow behaviour not only along the length of a dike, but also at different elevations. The NE flow directions of these samples indicate magma movement away from the volcanic centre, which is consistent with observations made by Shelley (1985).

The flow directions of samples 13628, 13619 and 13677, which indicate that magma adjacent to the country rock contact, moved away from the magma source, contrast with the flow direction of sample 13638, which indicates that magma adjacent to the upper contact at Castle Rock moved towards the magma source.

Samples 13628 and 13621 have flow directions which imply that throughout the trachyte (Groups A and B) emplacement, magma at the northern end of Castle Rock moved parallel to the magma source, back towards the volcanic centre (i.e. in a SW direction).

4.4.3 Benmoreites

The essentially horizontal magma movement indicated by the flow data implies that the surface over which the benmoreites flowed (i.e. the top of the Group B trachyte) was not inclined. The flow data for sample 13656 suggests that the emplacement of the high silica benmoreite was associated with SW magma flow. At the northern end of Castle Rock, low silica benmoreite magma movement was northward.

4.4.4 Comparison of flow data with experimental data

The flow directions established for the Castle Rock trachytes were compared with results from laboratory simulations of dome growth (Blake, 1990). These experiments produced lava domes by extruding kaolin slurry onto a horizontal surface from a central vent (Fig. 4.38). The central, elongate magma source implied by the Castle Rock configuration suggests that the experimental results are comparable in a general manner with the flow data from samples 13628, 13638, 13677, 13621 and 13619.

Two sets of intersecting surface textures, one radiating in a clockwise direction, the other radiating in an anticlockwise direction, were developed during the growth of these experimental domes (Fig. 4.39). Blake (1990) suggested that these could represent slip planes within the domes which facilitated magma movement during dome emplacement. From studies of movement directions defined by markers on the

surface of these experimental domes, Blake (1990) arrived at the following conclusions:

(a) New dome material only appeared at the dome surface on the dome summit.
(b) Surface material was not consumed by the expanding dome beyond the dome summit.
(c) Dome material was not overridden at the advancing dome front.

The flow directions of samples 13628, 13619 and 13677 adjacent to the trachyte lower contact are consistent with the observations made by Blake (1990), and imply that overriding of the trachyte at the advancing trachyte front did not occur. The magma movement towards the magma source, suggested by the flow direction of sample 13638, is not only at variance with the observations made by Blake (1990), but it is not the flow direction intuitively expected at the top of a growing lava dome. This may indicate that the application of Shelley's technique in determining flow directions at the tops of domes may not be valid. Alternatively, the elongate magma source inferred for Castle Rock and/or inhibited magma flow during the Castle Rock emplacement may preclude the analogy with Blake's experiments. If the methodology in determining the flow directions is valid, then it is postulated that the Castle Rock trachyte may have been intruded into shallow depths. When the lateral compression of the country rock exceeded the lateral tension of the magma and propagating fissure, magma intrusion was accommodated by upward movement of the

overburden. This forced magma flow at the extremities of the trachyte front to move back towards the magma source (Fig. 4.40 A, B). Restricted magma movement could also account for the essentially flat upper trachyte surface beneath the low silica benmoreite and the SW magma flow direction in the trachyte at the northern end of Castle Rock.

4.5 Groundmass crenulation fabric

4.5.1 Introduction

In thin sections cut parallel to the girdle axes, the groundmass feldspars of the trachytes and the benmoreites exhibit a crenulation (Fig. 3.16). The orientation and continuation of this crenulation is relatively constant across the thin sections, except adjacent to the phenocrysts and glomerocrysts, where it is deflected or arrested. Deformation textures, such as bending and breakage are not exhibited by the groundmass laths adjacent to the crenulation axial surfaces.

4.5.2 Relationship to flow data

The axial surfaces of the groundmass crenulation are orientated at a high angle to the flow lines. The relative movement displayed by the crenulations is consistent with that established by the analyses of phenocryst rotation

(Vernon, 1987) suggesting that the trachyte has experienced only one deformation episode.

The relationship of the crenulation axial surfaces to the margins of dike segment B1 suggests that the acute angle between them and the dike margins faces the magma source (Fig. 4.3, A). This observation is consistent with the results obtained by Shelley (1985) in a study of paleo-flow directions in radial dikes of the Lyttelton Volcano. The flow directions for the Castle Rock trachytes and benmoreites were established by assuming a similar relationship between the groundmass crenulation and the upper and lower contacts (Fig. 4.3, B).

4.6 Interpretation of magma emplacement

The platy shape was probably the controlling factor in orientating the groundmass and phenocryst feldspars in the trachytes (Groups A and B), high silica benmoreite and low silica benmoreite parallel to the flow line. The flow foliation produced by this strong preferred orientation was offset and rotated by, and developed prior to, the groundmass crenulation. The lack of deformation textures in the groundmass feldspars adjacent to or across the crenulation axial surfaces suggests that the crenulation of the groundmass must have occurred prior to solidification and before the laths had grown to their final size. The presence of intergrown feldspar laths across the crenulation axial surfaces implies that significant groundmass growth did occur

after the crenulation event. Given the present small amount of interstitial space between the laths it would be difficult to achieve crenulation of the groundmass without the development of deformation textures in the laths. Groundmass feldspar growth after groundmass crenulation was proposed by Shelley (1985) to account for similar groundmass textures in dikes of the Lyttelton Volcano.

The absence of brittle and plastic deformation textures in the phenocrysts and groundmass crystals in the Group B trachyte adjacent to the upper contact suggests that the low silica benmoreite was emplaced prior to trachyte solidification, but before groundmass growth had been completed. This evidence and the presence of coarse feldspar groundmass in the Group B trachyte adjacent to the low silica benmoreite relative to the smaller groundmass sizes in the rest of the trachyte, implies that significant cooling did not occur in the upper sections of the trachyte prior to the emplacement of the low silica benmoreite.

In a study of the paleo-flow directions in dikes of the Lyttelton Volcano, Shelley (1985) proposed that the obliquity of the groundmass fabrics to the dike walls was the result of velocity gradients established across the dikes during magma flow. The obliquity of the groundmass fabrics in the rocks of the study area suggests that velocity gradients were present at least within 0.5m of the dike segment B1 margins and the upper and lower contacts of the trachytes and benmoreites during fabric development. A velocity gradient across dike segment B1 may be reflected by the rotating flow foliation discernable in the field at The Tors locality (Fig. 2.10,

back pocket). The presence of crenulated groundmass in the trachytes and benmoreites distal from the contacts (e.g. sample 13682) suggests that these velocity gradients probably occurred throughout the magmas during emplacement.

The groundmass crenulation in the trachytes and benmoreites is reminiscent of the shear planes suggested by Blake (1990) to develop within growing domes to facilitate magma movement. It is reasonable to propose that the shear plane traces, developed on the surfaces of the laboratory simulated domes are the results of velocity gradients. If the groundmass crenulations in the trachytes and benmoreites of Castle Rock and the shear planes developed within the simulated domes are one and the same, it is postulated that magma movement during the emplacement of the trachytes and benmoreites within Castle Rock may have been facilitated by groundmass crenulation.

4.7 Synthesis

The strong preferred orientation of the feldspar groundmass and phenocrysts in the trachytes (Groups A and B), high silica benmoreite and low silica benmoreite was developed during magma flow. Velocity gradients during emplacement produced groundmass crenulations subsequent to the development of the flow foliation. These crenulations occurred in a magma - crystal mix where the high proportion of magma to crystals enabled strain to occur without brittle or plastic deformation of the crystals themselves. This

strain probably facilitated magma movement within the growing lava domes.

During trachyte emplacement, material was not overridden at the advancing magma front. Magma flow was predominantly away from the magma source but restricted magma movement at the northern end of Castle Rock and at the top of the trachyte may have produced magma movement back towards the magma source. This could reflect a shallow intrusion.

The emplacement of the trachytes was quickly followed by that of the benmoreites onto an essentially flat surface. Magma flow was away from the magma source and this produced westward and northward flow directions at the southern and northern ends of Castle Rock respectively. Feldspar preferred orientation did not develop in the basal sections of the low silica benmoreite.

5 Geochemistry

5.1 Introduction

In this chapter the major, trace element, isotope and REE analytical data on rocks from the Castle Rock area are presented. The geochemistry of the benmoreites (low silica and high silica) and trachytes (Groups A and B) within Castle Rock and the Eastern Sheet (hereafter referred to as the Castle Rock suite) is discussed in Section 5.2. The associated dikes are discussed separately.

Sample descriptions are presented in Appendix 1 and sample localities are found on Fig. 1.6. A review of analytical procedures is contained in Appendix 4 together with the major and trace element analyses, the normative mineralogy, isotopic and REE data.

Geochemical analyses were recalculated to 100% on a volatile - free basis prior to plotting. As discussed in Chapter 1, only samples with LOI < 2.5% are plotted on the variation diagrams.

The isotopic and REE data were obtained from S. D. Weaver. Some of these data have been published in a general account of Banks Peninsula geochemistry (Weaver & Smith, 1989), but the remainder are from unpublished work. Although these analyses were not obtained on samples collected by the author, a careful check on sample localities and whole rock geochemistry has established that they represent the trachytes (Groups A and B) and benmoreites (high silica and low silica) of the study area. Major and trace element

analyses are plotted against Zr, which was considered to be a useful "fractionation index." The XRF analytical method determines Zr with high precision and the Zr values in the rocks of the study area display a wide range of variation, particularly in the trachytes. During crystallisation of the major silicate phases in the benmoreites and trachytes, Zr behaved incompatibly and zircon was not found in the mode. "Volatile effects" during the late stages of crystallisation do not appear to affect Zr (Weaver et al, 1972) and Zr is generally regarded as immobile during subsequent weathering.

5.2 Major and trace element geochemistry

5.2.1 Introduction

Excluding sample 14027 from dike F, the rocks in the study area plot above the alkaline / sub - alkaline dividing line of Irvine and Barager (1971) indicating that they belong to the Alkaline Series (Fig. 5.1). A moderate Fe - enrichment, typical of alkaline volcanic series is exhibited by the rocks on an AFM diagram (Fig. 5.2) (Irvine & Barager, 1971).

Major and trace element variation trends suggest a fractionated volcanic suite consisting of benmoreites (low silica and high silica) and trachytes (Group A and Group B). Hawaiite and mugearite rock types are absent.

5.2.2 Major element variation

Variation with fractionation (judged by increasing Zr) occurs in all of the major oxides except Al_2O_3. With fractionation, TiO_2, MgO, CaO, Fe_2O_3, P_2O_5 and MnO decrease exponentially in contrast to the exponential increase of K_2O, Na_2O and SiO_2.

The highest rates of decrease or increase in the major oxides occur over the Zr interval 500ppm to 700ppm, which defines the transition from low silica benmoreite to Group B trachyte. A sharp inflection in these linear trends occurs at Zr values around 700ppm where the rate of decrease or increase in the major oxide variations is reduced dramatically or arrested.

The sharp increase in silica abundance (54wt% - 61wt%) exhibited by the low silica benmoreite to Group B trachyte trend is arrested at the Zr value 700ppm (Fig. 5.3). With increasing fractionation in the Group A trachyte (Zr 700ppm - 1380ppm) SiO_2 appears to be "buffered," its concentration being maintained between 61 - 63%. There appears to be a general increase in SiO_2 (61 - 63%) over the Zr range 700 - 1100ppm, followed by a slight decrease in SiO_2 at a Zr value around 1200ppm, however this latter variation may not be significant if a 2 sigma analytical error of ± 0.5% SiO_2 is considered.

The decreases in TiO2 (Fig. 5.4), MgO (Fig. 5.5) and Fe_2O_3 (Fig. 5.6) over the Zr interval 500 - 700ppm are accompanied by decreases in CaO (Fig. 5.7), MnO (Fig. 5.8) and P_2O_5 (Fig. 5.9) and contrasting increases in K_2O (Fig.

5.10) and Na_2O (Fig. 5.11). CaO, P_2O_5, K_2O, MgO, Fe_2O_3 and TiO_2 abundances remain relatively constant throughout the Group A trachyte.

The decreases in TiO_2, MgO and Fe_2O_3 may reflect the crystallisation of olivine, clinopyroxene and magnetite, whereas plagioclase and clinopyroxene crystallisation may explain the decrease in CaO. The decreases in P_2O_5 and MnO may be due to apatite and clinopyroxene crystallisation respectively.

The absence of a potassic crystallising phase in the low silica benmoreite to high silica benmoreite trend may be reflected by the increases in K_2O, however the smaller increase in Na_2O relative to K_2O over the same trend suggests that some of the Na_2O was removed by the crystallisation of plagioclase. The downward inflection of the Na_2O and K_2O trends in the trachytes suggests that these elements were removed by significant crystallisation of sodic and potassic phases. The scatter of data points displayed by Na_2O at the evolved end of the trend will be discussed in Section 5.2.4.

The constant Al_2O_3 (Fig. 5.12) throughout the low silica benmoreite to Group A trachyte trend reflects a bulk distribution coefficient close to 1.0. Thus Al - free and Al - poor phases (olivine and clinopyroxene) are approximately balanced by the Al - rich plagioclase.

These major element variation trends are reflected in the modal mineralogy. Olivine, clinopyroxene, plagioclase, magnetite and apatite are present in significant amounts in the benmoreites. In the trachytes, alkali feldspar is prevalent, calcic plagioclase is absent, olivine is rare, and

clinopyroxene, apatite and magnetite make up insignificant proportions of the mode.

5.2.3 Trace element variation

Variation with fractionation (judged by increasing Zr) occurs for all trace elements except Cr and Ni. Rb (Fig. 5.13), Nb (Fig. 5.14), La (Fig. 5.15), Ce (Fig. 5.16), Nd (Fig. 5.17), Th (Fig. 5.18), Y (Fig. 5.19) and Ga (Fig. 5.21) increase with fractionation suggesting that they were not concentrated in any of the major crystallising phases. Sr (Fig. 5.22) decreases exponentially throughout the low silica benmoreite to Group A trachyte trend and Ba (Fig. 5.23) initially increases linearly then decreases exponentially. Zn (Fig. 5.20) initially remains constant, until a Zr value of 700ppm, after which it increases linearly with fractionation. Many of these trace element graphs exhibit significant scatter along the trends and this will be discussed in Section 5.2.4.

As for the major oxides, the highest rates of decrease and increase in the Sr and Ba respectively occur over the Zr interval 500ppm - 700ppm which defines the transition from low silica benmoreite to Group B trachyte. These rates are reduced dramatically during fractionation of the Group A trachyte (Zr 700ppm - 1380ppm).

The Sr trend displayed by the benmoreites suggests that this element substituted for Ca during plagioclase crystallisation (Taylor, 1965). Sr may also substitute for K

in alkali feldspars (Wilson, 1989) and this may have occurred to a limited extent in the Group B trachyte.

The incompatible behaviour of Ba exhibited in the benmoreites suggests the absence of a major potassic crystallising phase and this is consistent with the evidence from the major element variation diagrams. The compatible trend displayed by Ba in the trachytes, indicates that it was substituting for K during crystallisation of the alkali feldspars (Taylor, 1965) particularly in the early formed Group B trachyte.

The absence of a potassic mineral phase in the benmoreites is also reflected by the incompatible behaviour of the Rb. The continuation of this incompatibility during fractionation of the Group B and less evolved Group A trachytes suggests that Rb was not substituting for K in the alkali feldspars. The incompatible and compatible behaviour of Rb and Ba respectively during fractionation of the trachytes suggests that the substitution of Ba for K during crystallisation of the alkali feldspars dominated over the substitution of Rb for K.

5.2.4 Dispersed trends

The data points on the Nb/Zr graph (Fig. 5.14) are concentrated along a well defined trend in contrast with the Sr (Fig. 5.22), Ba (Fig. 5.23) and Rb (Fig. 5.13) graphs, characterised by a few samples deviating from the well defined trends and the La (Fig. 5.15), Ce (Fig. 5.16), Nd

(Fig. 5.17) and Y (Fig. 5.19) graphs which have diffuse patterns of points along general trends. The data points on the Th (Fig. 5.18), Zn (Fig. 5.20) and Ga (Fig. 5.21) graphs produce trends which, although not as well defined as the Nb graph, are certainly not as diffuse as the La, Nd, Ce and Y graphs. The major element graphs produce well defined trends except Na_2O, which exhibits a dispersed trend in the Group A trachyte.

The dispersed trends on the Nd, Th, Zn and Ga graphs and the samples deviating from the trend on the Ba graph may not be significant if a 2 sigma analytical error for these trace elements is considered. However, the data points deviating from the trends on the Sr, Rb, La, Ce, Y and Na_2O graphs are outside this analytical error suggesting that volatile phases and/or zeolitisation and/or crustal contamination processes may have influenced the geochemistry of the rocks in the study area.

The weakly peralkaline samples in the Group A trachyte plot within the general trends of the Ce and La graphs suggesting that a late stage volatile phase during the crystallisation of the peralkaline magmas (Weaver et al, 1972, Weaver et al, 1990) was not responsible for the mobilisation of the Ce and La.

Studies on element mobility in Tertiary basalts (Wood et al, 1976) indicate that Si, Mg, K, Rb, Sr, and LREE can be mobilised during zeolitisation whereas Y and Nb are relatively unaffected. The variability percentage recorded by Wood et al (1976) in the remobilised elements covered a wide range, with Si and Mg displaying the lowest variability at ±

0.6% - 3.8% and ± 1.8% - 9.4% respectively.

The immobility during zeolitisation, implied by the coherent Si and Mg trends of the trachytes and benmoreites, suggests that the effects of zeolitisation on the rocks were minor. Mobilisation of La, Ce, Rb, Sr, Y, and Na during zeolitisation could be reflected in the dispersed trends on the graphs with the greater scatter, exhibited by La and Ce suggesting that the LREE were more readily mobilised. Na was mobile only in the trachytes, and this could be a reflection of the different crystallisation histories between the trachytes and benmoreites controlling the availability of Na during zeolitisation (Humphries et al, 1978). The increasing dispersion of this trend with fractionation suggests that Na was more mobile in the evolved Group A trachyte. The coherent Nd trend is consistent with the results above (Wood et al, 1976), however the dispersed trends exhibited by Y suggests that either it was more readily mobilised during zeolitisation relative to Y in the Tertiary basalts and/or other processes were affecting its distribution. Samples deviating markedly from the Sr and Rb trends contain amygdaloidal zeolites suggesting that a high degree of zeolitisation was required before these trace elements were mobilised. Dispersed trends due to crustal contamination effects will be discussed in Chapter 6.

5.3 Trachyte geochemical variation

The major and trace element variations in the Castle

Rock and Eastern Sheet trachytes are illustrated on Figs. 5.24 - 5.29. The least evolved trachyte, located adjacent to the upper contact in Castle Rock and including the trachyte dikes and sill cutting the low silica benmoreite, have been classed as Group B trachyte. The small geochemical variation within this trachyte contrasts with the wide geochemical range present in the more evolved rocks, which constitute Group A and include the trachyte from the middle and lower sections of Castle Rock, the Eastern Sheet and sill A. Samples from Area D (Fig. 1.6) fall predominantly at the evolved end of the Group A trachyte field, however the one sample (13694) from this area plotting in the Group B trachyte field (Fig. 5.24) is considered to represent the lower sections of the Castle Rock feeder dike.

Samples from the lower, middle and upper sections of Castle Rock define trends that span the entire geochemical range from Group A to Group B trachyte. Insufficient sampling due to inaccessibility on vertical sections could have produced the gap in the major (Fig. 5.24) and trace (Fig. 5.28) element trends between samples from the upper and middle sections of Castle Rock. Given that the same accessibility problem, present when sampling the lower sections of Castle Rock did not produce such a gap between samples from the middle and lower sections, it is considered that this gap could be real.

At a Zr value around 1000ppm, deviations from the linear trends of the Nb/Zr and Rb/Zr diagrams (Figs. 5.28, 5.27) are apparent and produce small, but significant decreases in the Rb/Zr and Nb/Zr ratios. The inflections of these incompatible

trends suggest that during the later stages of fractionation, a new mineral began to crystallise for which Nb and Rb partition coefficients were higher than Zr. However, the modal mineralogy of the Group A trachyte does not reveal such a phase. Alternatively, the decreasing Rb/Zr and Nb/Zr ratios in the Group A trachyte may indicate that in the more evolved rocks, Nb and Rb were incorporated into the existing crystallising phases.

5.4 Dikes

The geochemical relationship between the dikes and Castle Rock and the Eastern Sheet is illustrated on Figs. 5.3 - 5.23. The geochemical affinity of dike segments B1 and B2 to the most evolved Group A trachyte and the Group B trachyte respectively suggests that dike B was the feeder for the Castle Rock and Eastern Sheet trachyte. Likewise, geochemical similarities imply that dikes C and E could have been the feeders for the Castle Rock low silica benmoreite. The extension of dikes B and C into the intrusions was suggested by the occurrence of trachyte (Groups A and B) and low silica benmoreite in Area B and Group B trachyte in Area D (Fig. 1.6). Such evidence was not present for dike E, which terminated in outcrop metres before the intrusions (Fig. 2.1, back pocket), however the geochemical similarities between dikes C and E suggest that they were emplaced during the same event.

The slightly less evolved nature of the low silica

benmoreite dike A relative to the low silica benmoreite within Castle Rock suggests that it was emplaced subsequent to the Castle Rock intrusion.

The potassic (Fig. 1.9), sub - alkaline (Fig. 5.1) nature of dike F suggests that it is unrelated to the benmoreites and trachytes of Castle Rock.

5.5 Nb/Zr ratio

If the distribution coefficients between the crystalline phases and the liquid are equal during crystal fractionation, then the relative abundances of any two incompatible elements will remain constant (Gast, 1968). Weaver et al (1972) demonstrated on a Nb/Zr graph, that if this criteria is satisfied, rocks related by crystal fractionation and/or magma mixing from a common magma source will produce a straight line graph which passes through the origin.

A comagmatic origin for the low silica benmoreite, high silica benmoreite, Group B trachyte and the less evolved Group A trachyte from the middle sections of Castle Rock is indicated by the Nb/Zr graph (Fig. 5.14). The deviation of the more evolved Group A trachyte from the Nb/Zr trend is probably due to changes in the distribution coefficients between the liquid and the crystallising phases as the result of the evolved nature of the trachyte, and a different magma source is not inferred for the Group A trachyte.

5.6 Rare earth element geochemistry

The REE analyses for a Group A and Group B trachyte are presented in Appendix 4, Table 5.4. Chondrite normalised REE concentrations for the samples are similar (Fig. 5.30) and exhibit a strong LREE enrichment suggesting that the less evolved precursors of the trachytes (Group A and B) were derived either by small degrees of partial melting (Frey et al, 1978) or from a LREE - enriched source (Wilson, 1989). The REE enrichment, apparent in the Group A trachyte, suggests that REE abundances increased with fractionation. The negative Eu anomalies, displayed by both REE patterns confirm the more evolved nature of the Group A trachyte which has suffered more plagioclase fractionation than the Group B sample.

5.7 Spiderdiagrams

Normalised spiderdiagrams of Group A and Group B trachyte are illustrated on Fig. 5.31. From Rb to Ce and Tb to Yb the profiles are relatively parallel. The negative K anomaly apparent in both samples is similar to that found in other rocks on Banks Peninsula (Weaver & Smith, 1989). The presence of this anomaly in rocks from Canterbury, South Westland and the Dunedin Volcano, led Weaver and Smith (1989) to suggest that it may be a regional source feature. From Ce to Tb, the profiles are characterised by troughs at Sr, P and Ti, suggesting the removal of these elements by the

crystallisation of plagioclase, apatite and Fe - Ti oxides respectively (Thompson et al, 1982). The deepening of these troughs with fractionation reflects the increasing effect of these mineral phases on the more evolved rocks. The marked Ba trough exhibited by the Group A trachyte suggests the fractionation of alkali feldspar.

5.8 Sr - Nd isotope geochemistry

Sr isotope ratios from Castle Rock, dike segment B2 and dike C are presented in Appendix 4, Table 5.5, along with two Nd initial ratios from Castle Rock Group A trachyte and low silica benmoreite. Geochemical data pertaining to the samples are contained in Table 5.6 and sample localities are listed in Table 5.7.

The trachytes (Groups A and B), high silica benmoreite and low silica benmoreite are characterised by Sr initial ratios ranging from 0.70378 to 0.70304 with the highest and lowest ratios obtained from the Group A trachyte and the low silica benmoreite respectively. Sr initial ratios, intermediate between those of the Group A trachyte and low silica benmoreite, characterise the Group B trachyte (0.70336) and high silica benmoreite (0.70318). The Sr initial ratios of the trachytes (Groups A and B), high silica benmoreite and latterly emplaced low silica benmoreite suggest that these magmas were contaminated by continental crust (Chapter 6), although the Sr initial ratio (0.70323) of the low silica benmoreite in dike C may reflect Sr

mobilisation during zeolitisation (Section 5.2.4). The increase in Sr isotope initial ratios with fractionation indicates that contamination had a greater effect on the evolved rocks and this suggests that the low Sr magmas were more susceptible to contamination than the higher Sr (more mafic) ones.

The most evolved rock (Group A trachyte) and the least evolved rock (low silica benmoreite) have Nd initial ratios of 0.512672 and 0.512919 respectively. The positive Nd epsilon values imply that the Group A trachyte and low silica benmoreite were derived from a depleted mantle source (Chapter 6).

5.9 Conclusions

The relationship, between the trachytes (Groups A and B) and benmoreites (low silica and high silica), suggested by the major and trace element trends can be explained in terms of crystal fractionation, although magma mixing and/or crustal contamination processes can not be ruled out. Each of these processes will be modelled and discussed in Chapter 6.

Dikes B, C and E are considered to be the feeders for Castle Rock and the Eastern Sheet. Dike A is not associated with the intrusions and was emplaced after the Castle Rock low silica benmoreite. A late zeolitisation phase selectively remobilised elements, particularly in the trachytes.

Sr isotopic data suggest crustal contamination of the high silica benmoreite and trachyte magmas.

Isotopic and REE data indicate that the magmas were derived from a chemically related, depleted mantle source. However, the REE profiles suggest a LREE enriched source region. This conflict is resolved by considering a source which has a time - integrated depleted history but which has suffered enrichment (due to fluid or melt migration) in incompatible elements during or just prior to melting.

6. Petrogenesis

6.1 Introduction

In this chapter, constraints will be placed on the mantle source from which the mafic precursors of the benmoreites and trachytes were derived. In addition, crystal fractionation, crustal contamination and magma mixing processes will be quantitatively evaluated and discussed. Finally, geologic, petrographic, mineral chemistry, isotopic and geochemical data will be integrated into a petrogenetic model for the evolution of Castle Rock and the Eastern Sheet.

6.2 Mantle source

6.2.1 Introduction

The evolved nature of the benmoreites and trachytes precludes modelling of a specific mantle source, however recent studies have provided information on the general mantle source of Banks Peninsula mafic magmas (Sewell, 1985, Dorsey, 1988).

The benmoreites and trachytes of Castle Rock and the Eastern Sheet are considered to represent the evolved compositions of the alkaline olivine basalt series. It is generally accepted that basaltic magmas are derived by partial melting processes in the mantle and the parent mafic magma of the benmoreites and trachytes was probably derived

from a LREE enriched mantle source (Chapter 5) by partial melting (<15%) (Sewell, 1985) at pressures greater than 10kbar (Jacques & Green, 1980). The evolution of this partial melt by crystal fractionation and/or magma mixing and/or crustal contamination processes produced the low silica benmoreite, high silica benmoreite, Group B trachyte and Group A trachyte.

6.2.2 Data from Group A trachyte and low silica benmoreite

On a $^{87}Sr/^{86}Sr$ - $^{143}Nd/^{144}Nd$ plot (Fig. 6.1), the low silica benmoreite sample falls within the mantle array which suggests that it has not been contaminated by continental crust. The initial ratios of the Group A trachyte indicate that it has either been derived from a mantle source with lower Sm/Nd relative to that of the low silica benmoreite, i.e. less depleted, and/or contaminated by continental crust.

The depleted mantle source implied by the Sr and Nd initial ratios of the low silica benmoreite and Group A trachyte (Fig. 6.1) contradicts the LREE enriched mantle source suggested by the REE profiles. This apparent contradiction has been established for other Banks Peninsula lavas (Sewell, 1985, Dorsey, 1988, Weaver & Smith, 1989), and suggests that the mantle source was enriched in incompatible elements prior to or during partial melting (Chapter 5).

6.3 Geobarometry

The olivine phenocrysts and groundmasses of the benmoreites and trachytes have a high CaO content (>0.11wt%) suggesting that they were crystallised in a low pressure environment (Simkin & Smith, 1970).

Jadeite, Ca-Tschermak and titan-molecule components of selected diopside phenocrysts in the low silica benmoreite were calculated (Table 6.1) following the method of Munoz & Sagredo (1974). The titan-molecule ($CaTiAl_2O_6$) values (>1%) suggest crystallisation pressures considerably less than 10kb (Yagi & Onuma, 1967), whereas pressures less than 5kb are inferred from the low (<5%) jadeite ($NaAlSi_2O_6$) and Ca-Tschermak ($CaAl_2SiO_6$) values (Aoki & Kushiro, 1968, Munoz & Sagredo, 1974).

Early crystallisation of calcic plagioclase in an undersaturated magma at shallow depths can remove Ca-Tschermak molecule from the liquid resulting in clinopyroxenes which contain low amounts of CaO, TiO_2 and Al_2O_3 (Barberi et al, 1971). The presence of plagioclase inclusions in the diopside phenocrysts of the low silica benmoreite indicates that feldspar crystallisation commenced prior to clinopyroxene. However, the rare occurrence of these inclusions suggests that the role of these early crystallising plagioclases in determining the CaO content of the pyroxenes was not important. This hypothesis is supported by the absence of Ca-poor pyroxenes in the rocks of the study area and the presence of Al_2O_3 and TiO_2 values in the diopside phenocrysts typical of those found in alkaline basic rocks

(Fodor, 1975, Gibb, 1973).

The Fe - Ti oxide reaction rims on the kaersutite xenocrysts are inferred to be the breakdown products of high pressure hornblendes in a low pressure environment (Stewart, 1975). If a maximum crystallisation pressure of <5kb is inferred for the diopside phenocrysts, then the kaersutite xenocrysts probably crystallised at pressures greater than this. A high pressure environment is also inferred for the clinopyroxene phenocryst in sample 13656, which contains fresh kaersutite inclusions (Knutson & Green, 1975).

The presence of kaersutite xenocrysts suggests a hydrous, high pressure environment in the early stages of crystallisation whereas the crystallisation of plagioclase, clinopyroxene and olivine phenocrysts in the low silica benmoreite is associated with a low pressure environment (<5kb). Laboratory experiments conducted at low pressure indicate that plagioclase will commence crystallisation prior to clinopyroxene in the dry runs (Yoder & Tilley, 1962, Knutson & Green, 1975). The presence of plagioclase inclusions in the diopside phenocrysts of the low silica benmoreite suggests that anhydrous magma conditions may have been present during the later crystallisation stages.

6.4 Geothermometry

Crystallisation temperatures for selected plagioclase phenocrysts from the low silica benmoreite were calculated according to the method of Kudo and Weill (1970) with corrections applied after Mathez (1973). Microprobe analyses

of plagioclase phenocrysts in sample 13644 were used and bulk rock compositions were assumed to represent the liquids that were in equilibrium with the plagioclase liquidus temperature. This assumption was considered reasonable given the low phenocryst content (15% of the mode) of the low silica benmoreite. Lack of disequilibrium textures in the plagioclase phenocrysts indicates that these crystals were in equilibrium with the liquid. The crystallisation temperatures were calculated for the plagioclase phenocrysts with the highest and lowest An content.

The calculated temperatures over a range of PH_2O values (Table 6.2) range from 874°C at 0kb H_2O to 1291°C at 5.0kb H_2O. The anhydrous magma conditions thought to have been present during plagioclase crystallisation (Section 6.3) suggest that the temperatures calculated at 0kb - 0.5kb H_2O (874°C - 1039°C) are probably reasonable estimates for the low silica benmoreite. These crystallisation temperatures are similar to the crystallisation temperature (933°C) calculated for a ne-benmoreite from the Banks Peninsula Akaroa Volcanics (Dorsey, 1988).

6.5 Crystal fractionation

6.5.1 Introduction

A linear least squares approximation technique was used to quantitatively evaluate the crystal fractionation model suggested by the geochemical trends (Chapter 5). The results of the major element models were assessed by Rayleigh

fractionation modelling using published mineral/liquid distribution coefficients of selected trace elements.

The calculations were performed using the computer programme PETMIN (Eby, 1990), which is based on the method of Bryan et al (1969). All data were normalised to 100% volatile free and Fe was calculated as total FeO. MnO and P_2O_5 were excluded from the model.

The results were considered acceptable if the sums of the squares of the residuals were less than 1 and the calculated trace element abundances were within ± 20% of the observed abundances.

As the low silica benmoreite is the most "primitive" rock associated with the Castle Rock suite, it was assumed to be the magma from which the more fractionated rocks were derived. Tables 6.3 - 6.5 contain the distribution coefficients used in the calculations and the results of the least squares and Rayleigh fractionation modelling.

6.5.2 Results

The compositional variation between the benmoreites and trachytes can be achieved by a three step crystal fractionation model: (1) low silica benmoreite to high silica benmoreite, (2) high silica benmoreite to Group B trachyte, (3) Group B trachyte to Group A trachyte. The results of the major element modelling produce acceptable sums of the squares of the residuals at all three steps. Except for Ba at step 3, the calculated trace element results fall within ±

20% of the observed values. An unacceptably high distribution coefficient for Ba would have produced a better fit.

In steps 1 and 2, four mineral phases, plagioclase, clinopyroxene, olivine and magnetite, were subtracted from the proposed parent magmas to obtain calculated daughter liquids similar to the proposed daughter magmas. Plagioclase is the main fractionating phase in these two steps and the proportion subtracted increases from 66.3% in step 1 to 75.2% in step 2. The proportions subtracted for the minor fractionating phases, clinopyroxene, olivine and magnetite, decrease from 14.2%, 10.6% and 8.9% respectively in step 1 to 13.3%, 9.0% and 2.5% respectively in step 2. In step 3, alkali feldspar, clinopyroxene and magnetite, were subtracted from the proposed parent magma in proportions of 77.9%, 15.0% and 7.1% respectively. The relative proportions of the mineral phases subtracted from the parent magma at each step of the model compare favourably with the relative modal proportions of the phenocrysts. The general increase in mineral/liquid distribution coefficients with fractionation from low silica benmoreite to Group A trachyte is consistent with mineral-melt partitioning behaviour during increased polymerisation of a silicate melt (Le Roex & Erlank, 1982).

6.5.3 Conclusions

The results of the crystal fractionation modelling suggest that high silica benmoreite can be produced by crystallising 32% of the low silica benmoreite as

plagioclase, clinopyroxene, olivine and magnetite. Crystallisation of 23.8% of the high silica benmoreite using the same mineral phases, but in different proportions, can produce the Group B trachyte. Group A trachyte can be derived by crystal fractionating 31.3% of the Group B trachyte and subtracting alkali feldspar, clinopyroxene and magnetite. The inclusion of the mixed rocks (high silica benmoreite, Group B trachyte) in the model suggests that crystal fractionation was probably the dominant process in the magma reservoir. This model is not a unique solution, but it does suggest that crystal fractionation processes can produce the major and trace element geochemistry trends presented in Chapter 5.

6.6 Crustal contamination

6.6.1 Introduction

The crustal contamination suggested by the geochemical and isotopic data was quantitatively evaluated using the computer programme PETMIN (Eby, 1990). The two models generated, combined the effects of crustal contamination and crystal fractionation. Torlesse Supergroup, or their metamorphosed equivalents, was the assumed contaminant in both models.

In this section, the results of the modelling are presented and the effects of crustal contamination on isotopic, major, trace element and REE data are discussed.

6.6.2 Crustal contamination modelling

6.6.2.1 Model 1

The effect of crustal contamination on the benmoreites and trachytes was modelled using the method of De Paolo (1981). Sr initial ratios and geochemical data for the low silica benmoreite and Group A trachyte (Appendix 4) were obtained from S. D. Weaver. Some of these data were published in a general account on Banks Peninsula volcanism (Weaver & Smith 1989), the remainder are from unpublished work.

Torlesse, with Sr = 210ppm and $^{87}Sr/^{86}Sr$ = 710, was assumed to be the contaminant (Weaver, S. D. pers comm) and low silica benmoreite (sample 14, Table 5.7), with Sr = 705ppm (Table 5.6) and $^{87}Sr/^{86}Sr$ = 0.703 (Table 5.5), was used as the initial magma. The product of combined crystal fractionation and crustal assimilation processes was considered to be Group A trachyte (sample 62b, Table 5.7), with Sr = 11ppm (Table 5.4) and $^{87}Sr/^{86}Sr$ = 0.70378 (Table 5.5). Given the low percentage of phenocrysts in the mode of the low silica benmoreite (15%), an r (assimilation rate/crystallisation rate) of 0.01 was adopted. As plagioclase was a main fractionating phase, a bulk Sr K-D of 2.8 was considered reasonable.

The results of the model (Table 6.6) suggest that by crystal fractionating 90% of the low silica benmoreite and adding Torlesse, a daughter magma with similar Sr trace element and initial ratios as the Group A trachyte could be produced. The amount of crystallisation required to achieve

this relates well to the total crystal percentage (87.1%) derived from the three step crystal fractionation model (Tables 6.3 - 6.5).

6.6.2.2 Model 2

A linear least squares approximation technique was used to quantitatively evaluate a combined crystal fractionation - Torlesse contamination model. The mineral chemistry, whole rock analyses and mineral/liquid distribution coefficients of the three step crystal fractionation model (Section 6.5) were used, together with Torlesse geochemical data taken from Lobb (1988) (Table 6.7). Torlesse contaminant was added at each step of the model.

All data were normalised to 100% volatile free and Fe was calculated as total FeO. MnO and P_2O_5 were excluded from the model. The results were judged acceptable using similar criteria as those applied in the three step crystal fractionation model (Section 6.5).

The compositional variation of the benmoreites and trachytes can be achieved by a three step crystal fractionation model: (1) low silica benmoreite to high silica benmoreite, 2) high silica benmoreite to Group B trachyte, (3) Group B trachyte to Group A trachyte, with assimilated Torlesse (MA) of 4%, 2% and 2% at steps 1, 2 and 3 respectively (Tables 6.8 - 6.10). The major and trace element results are acceptable except Ba in step 3. The mineral phases and their proportions subtracted from the proposed

parent magma at each step are similar to those obtained in the three step crystal fractionation model (Tables 6.3 - 6.5), suggesting that crustal contamination was not a dominating process in the magma reservoir. The better major element fit in this model relative to that obtained in the three step crystal fractionation model is probably due to the added variable, Torlesse contaminant.

The greater amount of mass assimilated (MA) in step 1 may reflect the higher temperature of the benmoreites relative to the trachytes. Alternatively, if crystal fractionation processes are assumed to provide the heat required for assimilation (De Paolo, 1981) then the variable residence time of the benmoreites and trachytes in the magma reservoir may have influenced the amount of Torlesse assimilated by each rock type.

The relationship between the decreasing amounts of mass assimilated (MA) (4 - 2%) in the model and the decreasing Sr (766ppm - 14ppm) with fractionation suggests that smaller amounts of crustal contamination may have been required for the trachyte relative to the benmoreite, to produce the necessary change in the Sr isotopic ratio (Mantovani et al, 1985).

6.6.2.3 Conclusions

The results of the crustal contamination modelling suggest that the high silica benmoreite and the trachytes could have been contaminated by sediments of the Torlesse

Supergroup or their metamorphosed equivalents. The small amount of contamination indicated by the modelling and the low phenocryst content of the rocks implies that the benmoreites and the trachytes were erupted subsequent to a short residence time in the magma reservoir.

6.6.3 Effects of crustal contamination on isotopic, major,trace element and REE data

The benmoreites and trachytes are characterised by Sr initial ratios ranging from 0.70304 to 0.70378 with the least evolved rocks (low silica benmoreite) and the most evolved rocks (Group A trachyte) being the least enriched and the most enriched respectively in radiogenic Sr. The Sr initial ratio of the low silica benmoreite (0.70304) is typical of modern depleted mantle suggesting that these rocks have not been contaminated detectably by continental crust whereas the significantly increased Sr initial ratios in the high silica benmoreite and trachytes suggests that these magmas have suffered crustal contamination. The positive correlation between the Sr initial ratios of the benmoreites and trachytes and Zr (Fig 6.2) suggests that the evolved rocks were more susceptible to crustal contamination.

As Th and Rb are concentrated in the upper continental crust (Wilson, 1989), enrichment of these trace elements may reflect crustal contamination. Th in particular can be highly susceptible to contamination processes (Mantovani et al, 1985). To assess the Th concentrations in the benmoreites and

trachytes, a plot of Nb verses Th (Fig. 6.3) was produced following the procedure of Sewell (1985) and Dorsey (1988). The systematic decrease in the Nb/Th ratio suggests that the Group A trachyte may have been contaminated by continental crust and this hypothesis is supported by the Th enrichment in the Group A trachyte with respect to Ta and Hf (Appendix 4, Table 5.4).

Laboratory simulations of basalt contamination by continental crust (Watson, 1982) indicate that K may be the most mobile element during contamination processes, whereas Si and Na may be relatively unaffected. Watson (1982) concluded that the reason for selective contamination of K was its higher diffusivity relative to Si and reasoned that other elements e.g. REE, Nb and Zr could also be relatively mobile during crustal contamination. The absence of K enrichment in the Group A trachyte suggests that either crystal fractionation processes overprinted the effects of crustal contamination and/or that K was buffered against further change (Watson, 1982).

6.7 Magma mixing

6.7.1 Introduction

In this section, magma mixing processes suggested by the petrography and mineral chemistry are quantitatively evaluated. Discussions are presented on sieve texture development during magma mixing and the processes by which

magmas of different temperature and composition can be mixed. An analysis of sieve texture distribution in the Group B trachyte places constraints on the magma mixing event and the sequence of trachyte emplacement.

6.7.2 Model of magma mixing

A linear least squares approximation technique was used to quantitatively evaluate the combined effects of crystal fractionation and magma mixing. Crystal fractionation was taken as a one step process from low silica benmoreite to Group A trachyte and the mixed rocks were considered to be high silica benmoreite and Group B trachyte.

The calculations were preformed using the PETMIN computer programme (Eby, 1990), which did not include trace element modelling. All data were normalised to 100% volatile free and Fe was calculated as total FeO. MnO and P_2O_5 were excluded from the model.

The results were considered acceptable if the sums of the squares of the residuals were less than 1. Tables 6.11 and 6.12 contain the results of this major element modelling.

Major element fits were established for both mixed rock types, suggesting that combined crystal fractionation and magma mixing could have been operating in the magma reservoir. The crystal fractionation model of Section 6.3 included these mixed rocks and the major and trace element results of that model suggest that magma mixing was not the dominant process. This hypothesis is supported by the

scarcity of low silica benmoreite xenoliths in the mixed rocks and the absence of outcrop magma mixing criteria (e.g. streaking).

6.7.3 Magma mixing dynamics

Assuming that the groundmass feldspar laths grew during eruption and that the growth rates were uniform, then rims of similar width and composition should occur on the feldspar phenocrysts. The rims of the anorthoclase phenocrysts in the Group B trachyte are compositionally similar to the anorthoclase groundmass laths, however, the presence of phenocryst rim widths (0.06mm - 0.14mm) larger than half the width of the laths (0.03mm) indicates that crystallisation of the groundmass occurred subsequent to the onset of phenocryst rim growth. The absence of sieve texture in these phenocryst rims suggests that magma mixing was arrested prior to eruption.

Magma mixing as the result of convection in magma reservoirs has been established by experimental and theoretical studies (Huppert & Sparks, 1984, Huppert et al, 1982b, Sparks et al, 1984, Turner & Campbell, 1986). If, as the emplacement sequence of Castle Rock suggests, the magmas were erupted from a vertically zoned magma reservoir, then compositional and thermal gradients could have been established at the interface between the lighter, cooler trachyte and the underlying, denser, hotter benmoreite. These gradients may have been sufficient to force heat and chemical

components to move across the compositional interface, thereby establishing double - diffusive convection (Turner & Campbell, 1986) in the adjacent trachyte and benmoreite. The low phenocryst contents of the trachytes (1 - 3%) and benmoreites (10 - 15%) suggest that magma mixing due to convective fractionation (Sparks et al, 1984) may have been insignificant.

The small volume of mixed rocks (high silica benmoreite, Group B trachyte) relative to rocks exhibiting no evidence of mixing (low silica benmoreite, Group A trachyte) implies that magma mixing was not a dominant process in the magma reservoir. This hypothesis is supported by the results of the quantitative modelling and suggests that either the trachyte and benmoreite viscosities were too high to allow large scale mixing in the magma reservoir (Turner & Campbell, 1986) and/or that eruption of the magmas occurred soon after mixing commenced.

Many authors have suggested that magma mixing during the simultaneous withdrawal of magmas from a vertically zoned magma reservoir can occur during conduit flow (Blake & Ivey, 1968a, Koyaguchi, 1985, Blake & Campbell, 1986, Freundt & Tait, 1986, Blake & Ivey, 1986, Turner & Campbell, 1986, Koyaguchi, 1987, Koyaguchi & Blake, 1989). Laboratory simulations of such a scenario indicate that the behaviour of the two fluids (and the degree of mixing achieved) depends on their viscosities (Koyaguchi & Blake 1989). If these are large enough, then magma mixing could be initiated when the underlying less viscous fluid is drawn up through the overlying more viscous fluid (Koyaguchi & Blake, 1989). The

spatial and temporal relationship between the trachytes and benmoreites suggests that magma withdrawal was from the top of a vertically zoned magma reservoir and that in time magma was progressively drawn from deeper levels. However, the dikes and sill cutting the low silica benmoreite at the southern end of Castle Rock indicate that benmoreite was drawn up through and overtook a small proportion of Group B trachyte. If magma mixing was initiated during conduit flow, then the field and petrographic evidence implies that it must have occurred during the last stages of the eruption and this suggests that the trachyte and benmoreite viscosities were not large enough to initiate the instabilities required for magma mixing during conduit flow (Koyaguchi & Blake, 1989).

The degree of reaction experienced by crystals during a magma mixing event is dependent on the residence time of the crystals (Nixon, 1988) and the composition of the mixing magmas. If the completely sieved, partially sieved and unsieved resorbed anorthoclase phenocryst cores in the Group B trachyte are the result of different residence times and/or temperature and compositional differences between the mixing magmas, then the dominance of partially sieved phenocrysts at the northern end of Castle Rock suggests that these rocks were erupted either, soon after mixing commenced, or from areas of the magma chamber in which little mixing occurred (i.e. further from the mixing interface). This suggests that the trachyte at the northern end of Castle Rock could represent rocks emplaced earlier than those at the southern end.

6.8 Petrogenetic model for Castle Rock and the Eastern Sheet

The extensional tectonic regime, which persisted in New Zealand from the Late Cretaceous to the Middle Tertiary, triggered the Late Miocene volcanism on Banks Peninsula. This volcanism continued for 6 Ma and was associated with four main phases: Lyttelton, Mount Herbert, Akaroa, Diamond Harbour. Activity at the Lyttelton Volcano occurred during the earliest phase, and included the emplacement of Castle Rock and the Eastern Sheet during the late evolutionary stages of the volcano.

The following model incorporates geological, mineralogical, petrographical, geochemical and quantitative modelling data in an attempt to interpret the petrogenesis of Castle Rock and the Eastern Sheet.

Partial melting (<15%) of a LREE enriched mantle source at pressures greater that 10kb produced an alkaline olivine basalt magma which rose to a shallow depth in the continental crust (Fig. 6.4a). Crystal fractionation of this parent magma gave rise to a compositionally zoned magma reservoir in which the magmas became less evolved with depth. Presumably, the low silica benmoreite was underlain by more mafic magmas.

Prior to eruption, minor amounts of Torlesse Supergroup, or their metamorphosed equivalents, were assimilated into the magma reservoir and double - diffusive convection at compositional interfaces initiated magma mixing in the Group B trachyte and high silica benmoreite (Fig. 6.4b).

After a short residence time in the reservoir, the magma began to rise. As the pressure of this rising magma column

exceeded the lithostatic load, the conduit walls began to fracture and a vertical fissure propagated horizontally into the flanks of the volcano. This fissure tapped the top of the reservoir and withdrawal of magma commenced (Fig. 6.4c). In time, progressively deeper levels of the reservoir were tapped and the withdrawal of the Group A trachyte was quickly followed by the Group B trachyte, then the high silica benmoreite and latterly the low silica benmoreite. As the propagating fissure and intruding magma approached the volcano flanks, a decrease in the overburden compression accommodated emplacement of Castle Rock and the Eastern Sheet at shallow depths, close to or at the surface (Fig. 6.4d).

Another phase of conduit fracturing was associated with the final stages of magma emplacement and the subsequently emplaced low silica benmoreite was intruded via this new fissure. However, this renewed phase of fracturing was short lived and the withdrawal of magma was arrested prior to the complete emptying of the reservoir.

7. Conclusions

The mafic precursors of the benmoreites and trachytes were derived by partial melting of a mantle source, which suffered enrichment (due to fluid or melt migration) in incompatible elements during or just prior to melting.

Crystal fractionation of these magmas at shallow depths gave rise to a vertically zoned magma reservoir in which the Group A trachyte was underlain by the Group B trachyte, and the high silica benmoreite overlain by the latter. The low silica benmoreite magma occupied the deeper levels of the reservoir and presumably was underlain by more mafic magmas.

Crystal fractionation was the dominant petrogenetic process in this reservoir, however, isotopic and petrographic evidence indicate that crustal contamination and magma mixing processes did occur to a limited extent.

Torlesse Supergroup sediments, or their metamorphosed equivalents, are considered to be the contaminants assimilated into the reservoir during the evolution of the benmoreites and trachytes. The greater effect of this contamination on the more evolved rocks suggests that the trachytes were more susceptible to contamination than the benmoreites.

Magma mixing in the reservoir involved the high silica benmoreite and the Group B trachyte magmas and was probably initiated by double - diffusive convection at compositional interfaces. This mixing event was arrested prior to eruption, however a limited amount of magma mixing during conduit flow may have occurred in the later stages of intrusion.

After a short residence time, the magma was erupted when a magmatic stress regime, developed around a central conduit in the regular volcanic cone, initiated a fissure which tapped the top of the reservoir. In time, progressively deeper levels of the reservoir were tapped, however a small proportion of Group B trachyte and high silica benmoreite was emplaced after and/or during the emplacement of the low silica benmoreite.

As the propagating fissure and intruding magma approached the volcano flanks, a decrease in the overburden compression accommodated horizontal and vertical expansion of the fissure and enabled Castle Rock and the Eastern Sheet to be emplaced at shallow depths close to or at the surface. Withdrawal of magma was arrested prior to the complete emptying of the reservoir.

During magma flow, the feldspar phenocrysts and groundmasses of the trachytes and benmoreites acquired a strong preferred orientation, which was offset and rotated by, and developed prior to a groundmass crenulation. These crenulations occurred in a magma - crystal mix where the high proportion of magma to crystals enabled strain to occur without brittle or plastic deformation of the crystals themselves. This strain probably facilitated magma movement within the growing dome.

Magma was not overridden at the advancing magma front during the emplacement of Castle Rock. The trachyte magma flow was generally away from the magma source, however, restricted magma movement at several localities may have been responsible for magma movement back towards the source. The

emplacement of the trachytes was followed by that of the benmoreites onto an essentially flat surface.

Subsequent to the emplacement of Castle Rock and the Eastern Sheet, a late zeolitisation phase selectively remobilised elements particularly in the trachytes.

Castle Rock and the Eastern Sheet were emplaced during the late evolutionary stages of the Lyttelton Volcano, Banks Peninsula. This Late Miocene volcanic activity represents intraplate volcanism within an extensional tectonic regime.

Acknowledgements

The completion of this project would not have been possible without the assistance and support of many people including the staff, technicians and students at the Universities of Canterbury and Victoria, friends, relatives and family.

To my supervisor Steve Weaver, I would like to express my thanks and appreciation, for his encouragement to undertake the project, his tireless enthusiasm throughout the duration of the project, his patience in editing numerous pages of text, his advice on geological and non geological matters and in particular for his support when the odd MSc crisis occurred. David Shelley taught me the intricate U - stage techniques and took time out from sabbatical leave to read text. Jim Cole assisted me when Steve was in Antarctica and acted as a buffer between the University system and myself.

The expertise and support of the Canterbury Geology Department technicians was invaluable during the course of this project. Countless thin sections (including orientated thin sections), polished mounts and polished sections were produced by Karen Holder. Steve Brown provided XRF expertise and together with Katherine unravelled the major problems encountered with the Word Perfect and PETMIN computer programmes. Valuable advice on photography techniques was given by Albert Downing, who patiently went through the procedure for photographing thin sections a second time when the first three films failed. Arthur Nicholas taught me the correct maintenance and use of the rock cutting saws and Lee Leonard provided me with advice on draughting techniques. To Joan and Julie Anne, well watch out, I am now so well versed in Word Perfect 5.1 I could take over your jobs.

The assistance and support given by students (post graduate and under graduate) at the Canterbury Geology Department is gratefully acknowledged. Andy Nicol taught me the deeper mysteries of Word Perfect 5.1, read numerous pages of text on fabric work, provided me with ideas on structure and taught me how to survive an MSc intact. Roddie Muir assisted in rotating orientated samples, read pages of text and together with Karen assured me that one day the project would be finished. Pages of lists, references, captions (all the boring bits) were edited by Barbara Hobden (who never complained) and Todd Warner assisted in sample collecting and unravelled the hiccups encountered in PETMIN. Thank you Viv for not commenting about the bedlam in the room during the last few months, for stimulating conversations over the last two years, for your sense of humour which provided me with light relief when everything got too serious, for not mentioning that I had failed to return your mini trampoline (after six months) and for putting up with my lack of conversation in the last few weeks. To Ritva Karhunen an extra special thanks for not saying anything about the chaos in the flat during the last two months.

I would to thank the staff and students at Victoria University who made my stay so welcome and to Jo and Co and Kath and Co for providing me with a bed whilst in the capital. Paul, Richard, Phil and Rodney taught me how to get information out of the microprobe and Ken patiently fixed up my mistakes without complaining.

The support of my family over the last fifteen months has often provided me with the impetus to continue with this project. I would to thank my brother Patrick, for the loan of field equipment and for not commenting when twelve months later it was still not returned; Julie, for continued interest in my progress; Anne and Grant, for supplying me with numerous meals particularly in the last few months; Helen, for many hours of cutting, pasting and colouring; Rob, for preparing me for the MSc merry - go - round. And lastly, but no less importantly, I would like to thank my father Pat, who has encouraged me throughout and in particular for teaching me that anything is possible.

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